CIG XIX IGC

CONGRÈS INTERNATIONAL DE GÉOGRAPHIE

NORDEN 1960

INTERNATIONAL GEOGRAPHICAL CONGRESS

Glacial Morphology and Inland Ice Recession in Northern Sweden

by GUNNAR HOPPE

The Glaciers of the Kebnekajse-massif

by VALTER SCHYTT

Appendix: Ice Melting under a Thin Layer of Moraine, and the Existence of Ice Cores in Moraine Ridges

by GUNNAR ÖSTREM

The Inland Ice in the Abisko Area

by C. G. HOLDAR

The Kårsa Glacier and its Relation to the Climate of the Torne Träsk Region

by C. C. WALLÉN

EXCURSIONS E. SW. 1, E. SW. 2 A, E. SW. 2 B

GLACIAL MORPHOLOGY AND INLAND ICE RECESSION IN NORTHERN SWEDEN

BY GUNNAR HOPPE

Department of Geography, University of Stockholm

INTRODUCTION

The primary aim of this paper is to attempt to give a general view of the final stage of the Würm (Wisconsin) glaciation in Northern Sweden and to demonstrate the principles of deglaciation, late changes in the directions of ice movement, and the morphological activity of the ice remnants and their meltwater streams. The following discussion mainly concerns the province of Norrbotten, the northernmost fourth of Sweden.

The bedrock of eastern Norrbotten is of Archaean age. It belongs to two different cycles, the Svionian, which is believed to have ended about 1,800 million years ago, and the Karelian. The age of the Karelian cycle and its relation to the Svionian is at present an intensely debated subject. Granites and gneisses dominate but there is also a great variety of other rocks: volcanics and sediments, for instance, play an important role (see Ödman 1957). In western Norrbotten the Archaean is overlain by younger, mainly metamorphic, rocks belonging to the Caledonian Mountains. Between the Archaean and the metamorphic Caledonian areas there is in most places a narrow strip of autochtonous Eocambrian and Cambrian rocks.

Along the Gulf of Bothnia and inland to a breadth of up to 30 or 40 km. runs a coastal plain on which differences of elevation seldom exceed 50 meters. The coastal plain slowly gives way to a terrain with hills which may rise 200 to 300 meters above the surrounding larger or smaller plains. This area has been characterized as an inland plain with residual hills—monadnocks; similarities between this area and the "Inselberg" landscapes of the semiarid tropics have been previously pointed out (for discussion see Rudberg 1954). To the west of the inland plain lies a premountain area and then the true mountain region, where elevation differences of more than 400 meters are frequent. This region includes the biggest high-mountain complexes of Sweden (Kebnekajse, Sarek, etc.) with several peaks of more than 2,000 meters. Through the mountain chain run a number of river valleys in a NW—SE direction; in all of them elongated lakes, often quite deep (Hornavan 221 meters), appear; so it is justified to speak of a *mountain and lake region.* The morphology of the mountains shows the effects of "nappe" tectonic forces which have produced steep slopes at the overthrust boundaries, most of them oriented toward the east.

The surface of the bedrock of Norrbotten as well as in other parts of northern Sweden is considered by many scientists to form a "Piedmonttreppe". The most important study along these lines is that of Rudberg (1954) in his investigations in Västerbotten, the province lying to the south of Norrbotten. He is there able to distinguish not less than 13 erosion cycles.

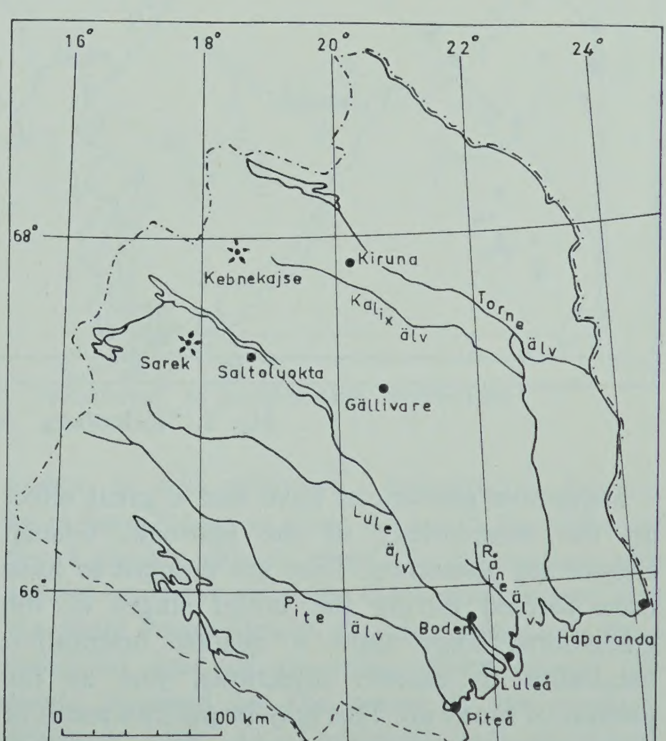


Fig. 1. Location map.

1—910100 Sep.

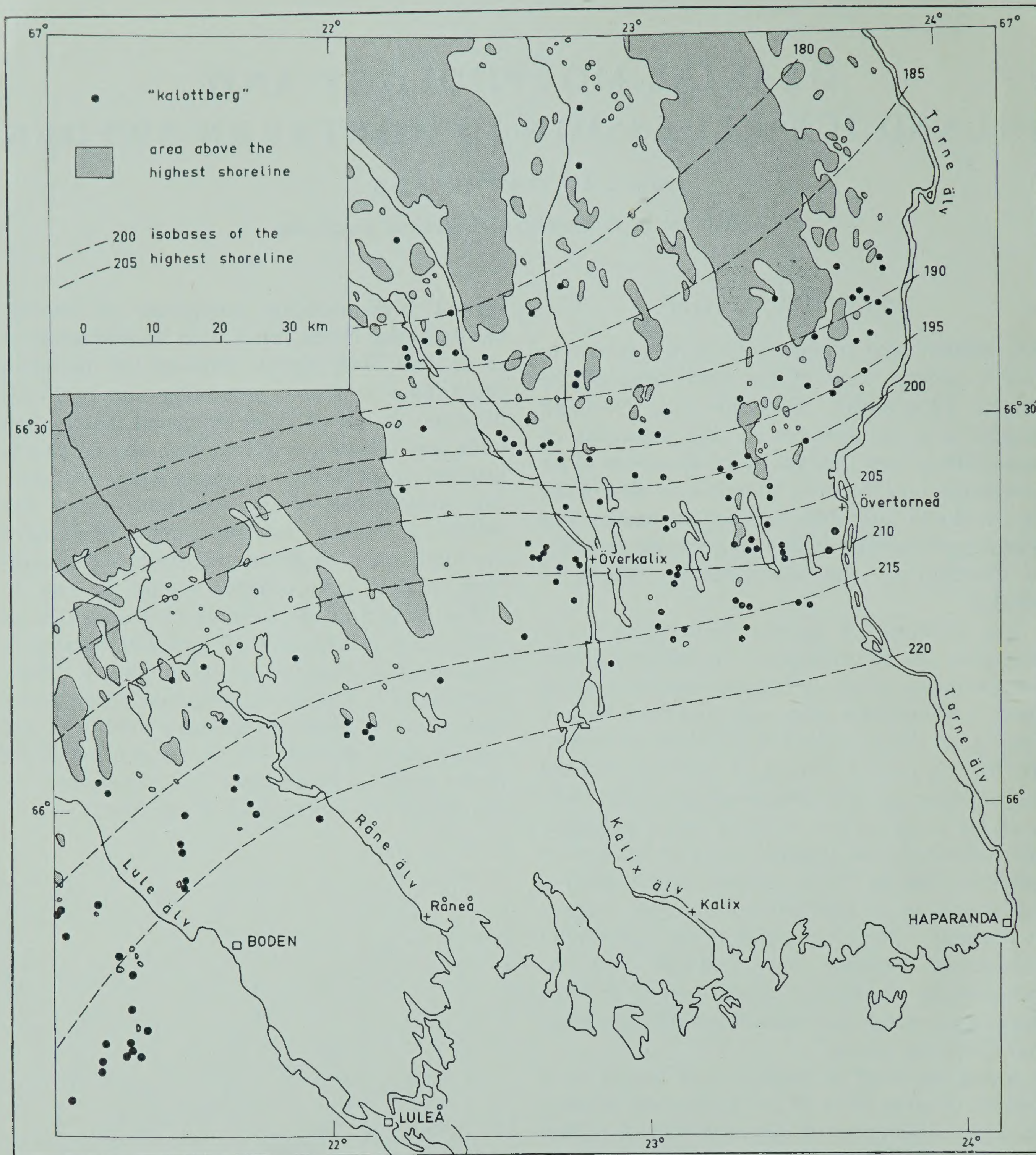


Fig. 2. "Kalottberg" in southeastern Norrbotten.

Successive glaciations have had a great effect on the morphology of the bedrock. Glacial cirques are numerous. They are thought to have been formed during the initial stages of the glaciations. They show a definite orientation maximum to eastern directions, just as the glaciers of today do. This may be partly a result of the overthrust steps, but the main reason presumably is the dominance of winds from the west, which have deposited snow especially on the leeward sides. The lowest cirques in Norrbotten appear on hills with a height of 800—900 meters (Ljungner 1949, p. 38; own observations), which means a depression of the glaciation limit (or firn limit) of about 1,100 meters as compared with the present one. This, however, is not necessarily the same as the maximum depression, since the whole area was inundated by ice at an

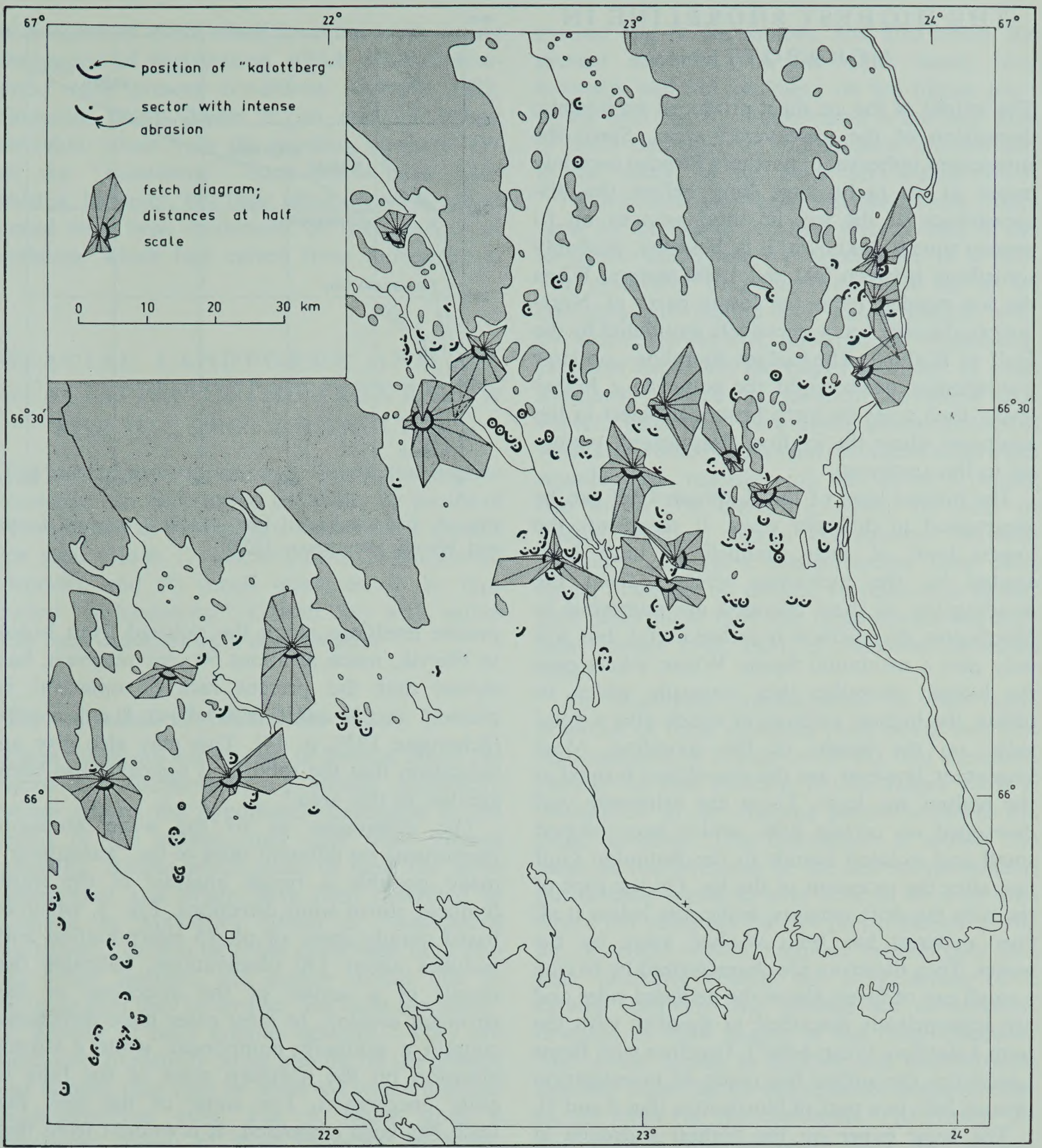


Fig. 3. Direction of intense abrasion on the "kalottberg" in southeastern Norrbotten.

early stage.—The valleys, especially those oriented in the direction of the ice movement, have been transformed into glacial troughs, most typically in the mountain area, but also existing everywhere in topographically less pronounced terrain. Hills and mountains quite often have the appearance of gigantic *roches moutonnées* and drumlins.

The number of successive glaciations in Norrbotten is still unknown. Interglacial deposits have been found in some places in the valley of the Lule älv (älv = 'river') in connection with studies undertaken by the Geological Survey and the State Power Board. C_{14} determinations have indicated ages of an order which exceed the limits imposed by the range of the method. (Fromm 1960, G. Lundqvist oral communication.)

THE HIGHEST SHORELINE IN NORRBOTTEN

The weight of the ice sheet produced an isostatic depression of the ice-covered areas. Since the subsequent upheaval in northern Sweden certainly began at an early stage long before the disappearance of the ice, its total amount up to present time is unknown; it is, however, probably something between 500 and 1,000 meters. When the ice receded from the lower parts of Norrbotten, the area was successively inundated by the Gulf of Bothnia. The highest shoreline—no later transgression has brought the water to a higher level—then must be metachronous, oldest in the southeast where the ice first disappeared, youngest in the northwest.

The highest level of the Bothnian Gulf can be ascertained in different ways. It represents the lowest limit of such channels as have been eroded by the meltwater streams from the receding ice; as these channels are numerous in Norrbotten this method is rather useful, but will only give a maximum figure. Where eskers pass the highest shoreline they normally widen to deltas, the highest surfaces of which give a good value on the height of the shoreline. Most important, however, are the strandlines formed at the highest sea level. These are extremely well developed on certain hills, which once formed small and isolated islands in the Bothnian Gulf just after the recession of the ice. On the tops of the hills the drift remains, while just below it all finer material has been washed away by the waves. They therefore are characterized by having a small cap of forest above the denuded sides and are appropriately described in Swedish with the term *kalottberg* ("cap-hills"). Together with Bernt Lindström the author has made an investigation of such hills in a part of Norrbotten (figs 2 and 3).

The basic paper on the highest shoreline in Norrbotten is that of Santesson (1927); more recent investigations (Hoppe 1948, p. 67; Fromm 1949, p. 313, and oral communication) on the whole have only confirmed the data given by him. This shoreline is highest in the neighbourhood of the present coast, about or slightly exceeding 220 meters above sea level, then descends toward the north and northwest, down to values of about 170 meters. The differences are due to the different time of deglaciation as mentioned above, but probably also to a different rate of land upheaval. A comparison (fig. 4) between two

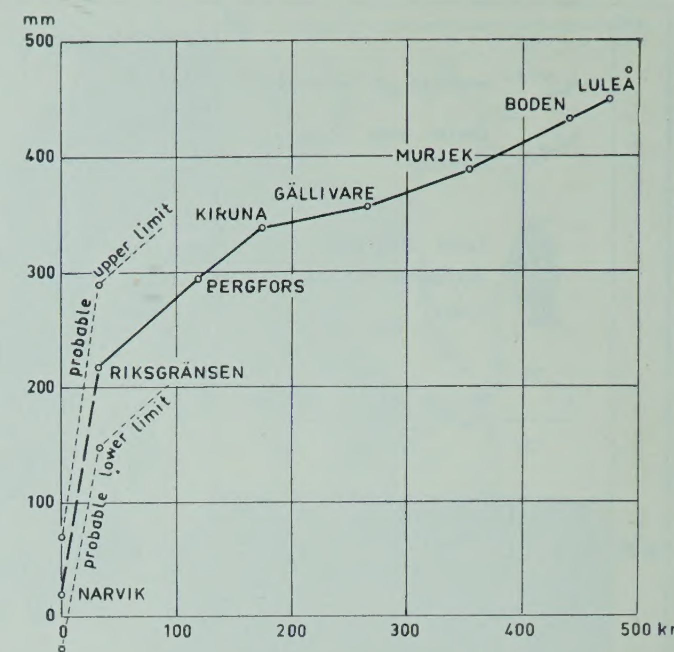


Fig. 4. Land upheaval 1906—1956/57 between Luleå and Narvik (SCHERMAN 1959).

precise levellings along the railroad from Luleå to Narvik, made at about 50 year intervals, has shown that the present rate of upheaval is greatest at the coast with about 1 cm. a year (Scherman 1959, p. 14). This may also give an indication that the total land upheaval has been greatest in this area.[1]

The differences as to the wave abrasion phenomena on different sides of the "kalottberg" make possible a rough analysis of the most frequent storm wind directions. Fig. 3, which is based mainly upon air photo interpretation and includes about 130 observations, indicates the results of a survey of the directions of the strongest erosion. In most cases these directions contain a southerly component, while a strong abrasion on the northern sides of the hills is quite uncommon. For some of the hills the fetch has been computed. It is evident from this map, that the abrasion is normally well correlated with the fetch, and this is one and probably the main reason, why a strong abrasion from the north only seldom appears. It may be observed, however, that even in those cases where the fetch is about the same from northerly and southerly directions, the stronger abrasion is on the southern side. This may have been due to a

[1] In a recent paper Hast (1959, p. 29), however, asserts that the present upheaval is not an isostatic readjustment due to the depression of the last ice-sheet. His opinion is based on measurements of the rock pressure in mines.

dominance of storm winds from the south during the period of deglaciation, which would correspond with present conditions (Östman 1922, Enequist 1944). There is no sign of strong catabatic winds from the receding ice-sheet NW of the "kalottberg". Such winds may have existed, however, but their effect on wave action could have been moderated by the presence of icebergs, which had calved from the ice-front.

GLACIAL LANDFORMS AND DEGLACIATION IN THE AREA BELOW THE HIGHEST SHORELINE

The deglaciation in the area below the highest shoreline can be followed especially by means of (1) moraines of the type first described by Gerard De Geer (1889), (2) glacial striae, (3) glacifluvial deposits, and (4) thrust moraines of the type called "Kalixpinnmo". Glacial clay with varves has been observed only occasionally.

The moraines of Gerard De Geer

These ridges are well known in Sweden under the name of "årsmoräner" (= annual moraines), which reflects the hypothesis—untenable in the mind of the author—that they were formed annually. Ridges which seem to have the same appearance and origin have been described in Canada by Mawdsley (1936) as "washboard moraines". As this term has since been used also for other features, the author would like to propose that they be termed "De Geer moraines" in honour of their famous detector.—They are quite easy to identify on air photos; on the present sea bottom they may be observed where detailed echo-soundings exist. Fig. 5, which gives a schematic survey of these moraines in Norrbotten, is based on air photo interpretation, field studies and—for small areas—thorough investigations of echo sounding material.

The De Geer moraines in Norrbotten have a height of up to 6 or 7 meters, a breadth of from 8 to 40 meters and may attain a length of one km. In cross-profile they are often oblique with a steeper distal (= down-glacier) side having a greater number of loose boulders. They normally consist of till, sometimes with pressure structure or, rarely, sedimentary lenses. They only appear below the highest shoreline, especially on level ground and in the valleys, and also below the present sea-level. On the other hand, they normally will not be found on the higher parts of the hills (fig. 6). Unexplained is the total lack of such moraines in southeastern Norrbotten, in the valleys of the Kalix and Torne rivers.

All scientists who have discussed these ridges agree to the extent that they were formed at the ice-front, which had the appearance of a cliff. The possibility that they were built up in ice crevasses, just behind and mainly parallel with the front, should not, however, be overlooked. De Geer explained the ridges as a sort of push moraines, formed by ice-front advances each winter; thus they were given geochronological value as the distance between two succeeding ridges equalled the recession of the ice-front during one winter (De Geer, e.g. 1940). In areas in middle Sweden, where comparisons between moraines and varve measurements have been made, however, no definite correlation is found; very often more than one ridge was formed each year (Hoppe 1948, 1957; cf., however, Järnefors 1956).

An important consideration in the interpretation of these moraines must be the fact first demonstrated by Lundqvist (1948), that elongated pebbles in free position in the ridges show a disposition to be oriented perpendicular to the ridge direction. This orientation is usually the same as the direction of the last ice-movement, but actually seems to have little to do with this; where the moraines are not perpendicular to the ice direction—which happens occasionally and then probably reflects the irregularities of a calving ice-front—the pebbles are nevertheless oriented at right angles to the ridge direction. This would appear to indicate that water-soaked moraine under the ice was squeezed perpendicular to the front (or the crevasses). The reason for this motion was probably that the seepage pressure rose above a critical value under the influence of changes in a water-filled system and/or of the calving mechanism (for details, see Hoppe 1957). The author admits that the above explanation may not cover all possibilities and may require further elaboration.

The De Geer moraines as ice-front features demonstrate some characteristics of the ice retreat. Thus in areas with a broken topography, the front receded faster where the water depth was greater, a behaviour which we can observe in glaciated regions today, (e.g. Muir Inlet, Alaska;

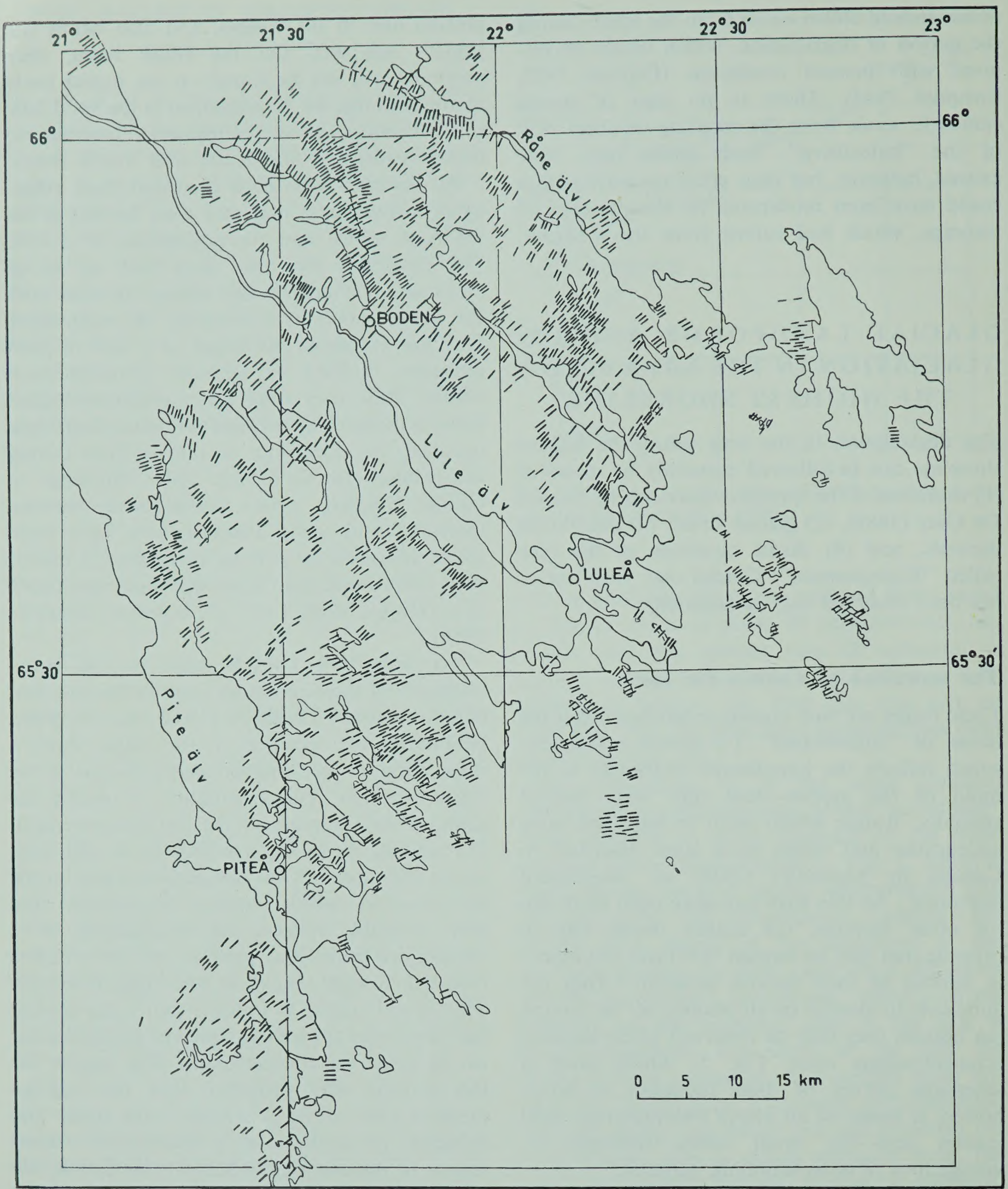


Fig. 5. Schematic map of DE GEER moraines in Norrbotten. (From HOPPE 1948, with some complements).

Field 1947). So-called calving-bays (fig. 6) were then formed, the term emphasizing the importance of the intensified iceberg calving due to a stronger uplift; on the other hand the melting may also have been stronger in the same places, because of differences in water circulation and temperature.

Glacial striae

In areas which are situated only some tens of meters above the present sea-level, the *roches-moutonnées* usually show a distinct glacial striation. On higher levels much of the scratching has disappeared especially on more coarse rocks.

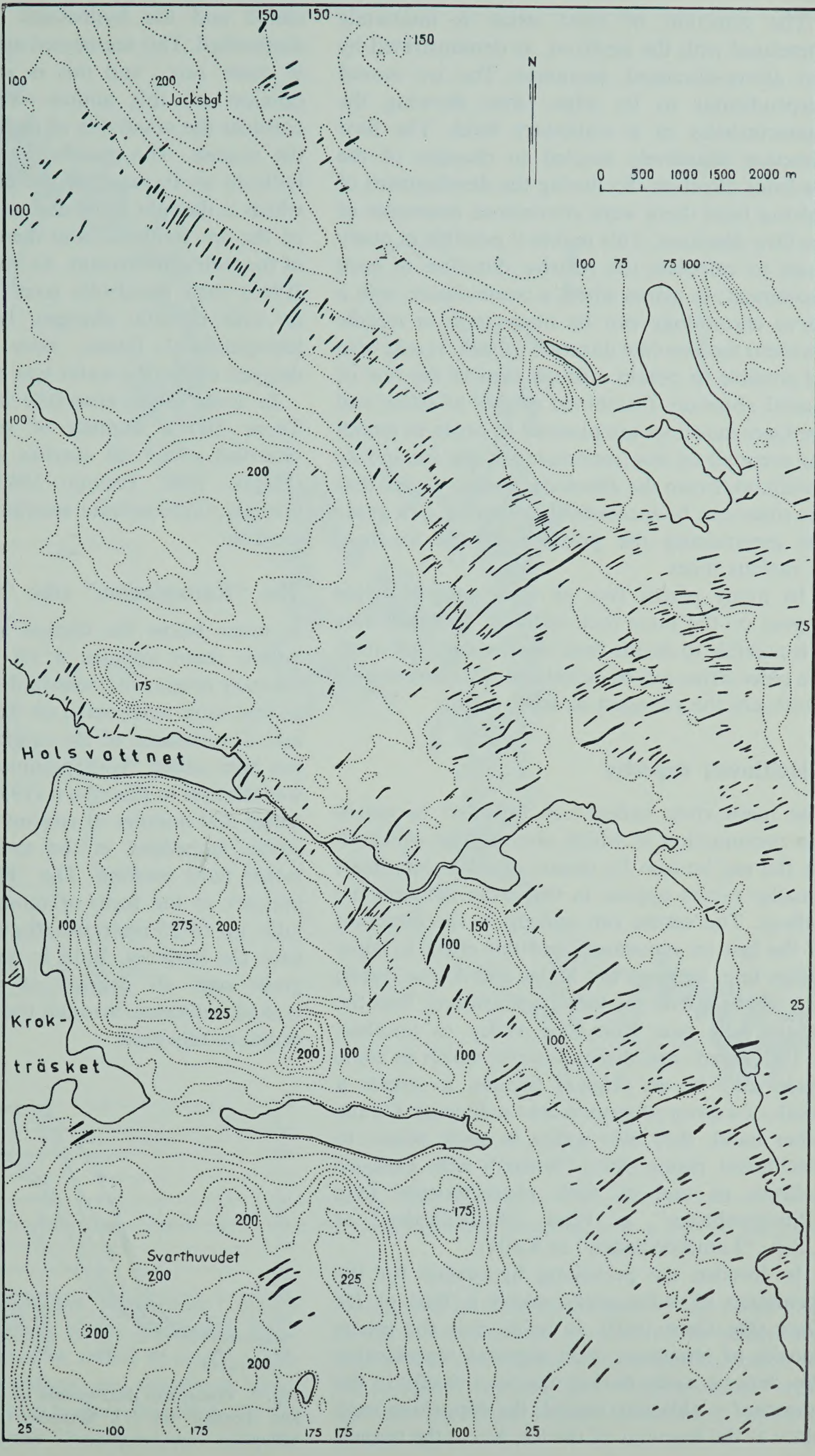


Fig. 6. De Geer moraines in an area north of Boden (Hoppe 1948). The moraines are situated mainly in the valleys and show the development of calving bays. Contour interval 25 meters.

The direction of most striae is intimately correlated with the ice-front, as demonstrated by the above-discussed moraines. The ice moved perpendicular to its edge, thus showing the characteristics of a stationary fluid. The flow direction sensitively reacted to changes of the ice-front direction. So during the development of calving bays there were continuous responses of the flow direction. This makes it possible in many cases to calculate the relative duration of each movement; in places where a combination with a varve chronology can be made (e.g. in middle Sweden) the absolute duration as well. It may also be possible to obtain a rough idea of the rate of glacial abrasion. Combined studies of striae and De Geer moraines are planned in order to search for eventual *en bloc* movement of the ice and its condition. From the above discussion we see that the striae can be successfully employed as a guide for ascertaining the position of the ice-front at various times.

In many places two or three sets of striae appear on the same rock outcrop. In most cases it is possible to decide their relative age, and often the older striae can be correlated with movements which are the youngest in other areas.

Glacifluvial deposits

The main river valleys are followed by eskers the continuation of which also may be observed on the sea bottom by means of echo-soundings. Smaller eskers appear in the areas between the valleys. The eskers run mainly in the direction of the last ice movement and this effect in some places (e.g. between the Kalix and Torne rivers) was stronger than the topographical one, thus the eskers may pass from one valley to another.

The eskers are discontinuous, which in some areas may be an original feature, in others a result of a cover of more recent sediments. On the other hand, they also widen in some places to glacifluvial plains (Sw.: "hedar") with dead-ice hollows as one of their characteristics (e.g. "Pitholmsheden" at Piteå, "Kallaxheden" at Luleå, "Lantjärvsheden" at Kalix).

In Sweden the prevailing hypothesis for the formation of subaqueous eskers is that of De Geer (De Geer 1940). It holds that the eskers consist of segments, each segment representing the deposit made during one year chiefly in the summer (= ablation) period; the deposition took place at the terminus of the ice, where the tunnels ended and the hydrostatic pressure suddenly diminished. This segmented appearance, however, is quite rare, and this is one reason—in the opinion of this author—why one must also consider the possibility of deposition while still in the tunnels. The glacifluvial plains ("hedarna") indicate an increase of the deposition per area, which is thought to be due to a relative standstill of the ice terminus, and then also of the mouth of the meltwater stream. As discussed by Enequist (1946) such standstills seem to have nothing to do with climatic changes, but were caused by topographical forces, since the calving was delayed where the water was shallower.

In some areas, especially in the valley of the Torne älv, a number of sections have been observed where till overlies glacifluvial deposits (Hoppe 1948, Fromm 1949). These sections probably demonstrate smaller oscillations of the ice front.

The "Kalixpinnmo" (the "Kalix till")

In areas below the highest shore-line in Norrbotten there appears a till of evidently sedimentary origin, characterized by a predominance of fine sand and coarse silt (0.2—0.02 mm, fig. 7) and disturbances of the original stratification. It has been called "Kalixpinnmo" by Beskow, who was the first to describe it (1935). Typical also are lenses and nodules of coarser sediments, angular stones embedded in the sand and silt, and a rather hard packing. The "Kalixpinnmo" often appears in the form of more or less elongated hills, up to 25 meters high and oriented parallel with the ice-front. Both in terms of texture and grain size, all degrees of transition between typical sediments through the "Kalixpinnmo" to ordinary tills exist.

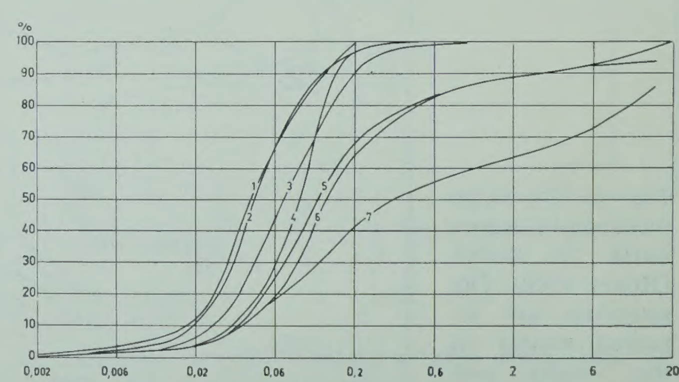


Fig. 7. Grain size distribution of «Kalixpinnmo» (1—6) and normal till (7). Samples from Kalix and Råne valleys.

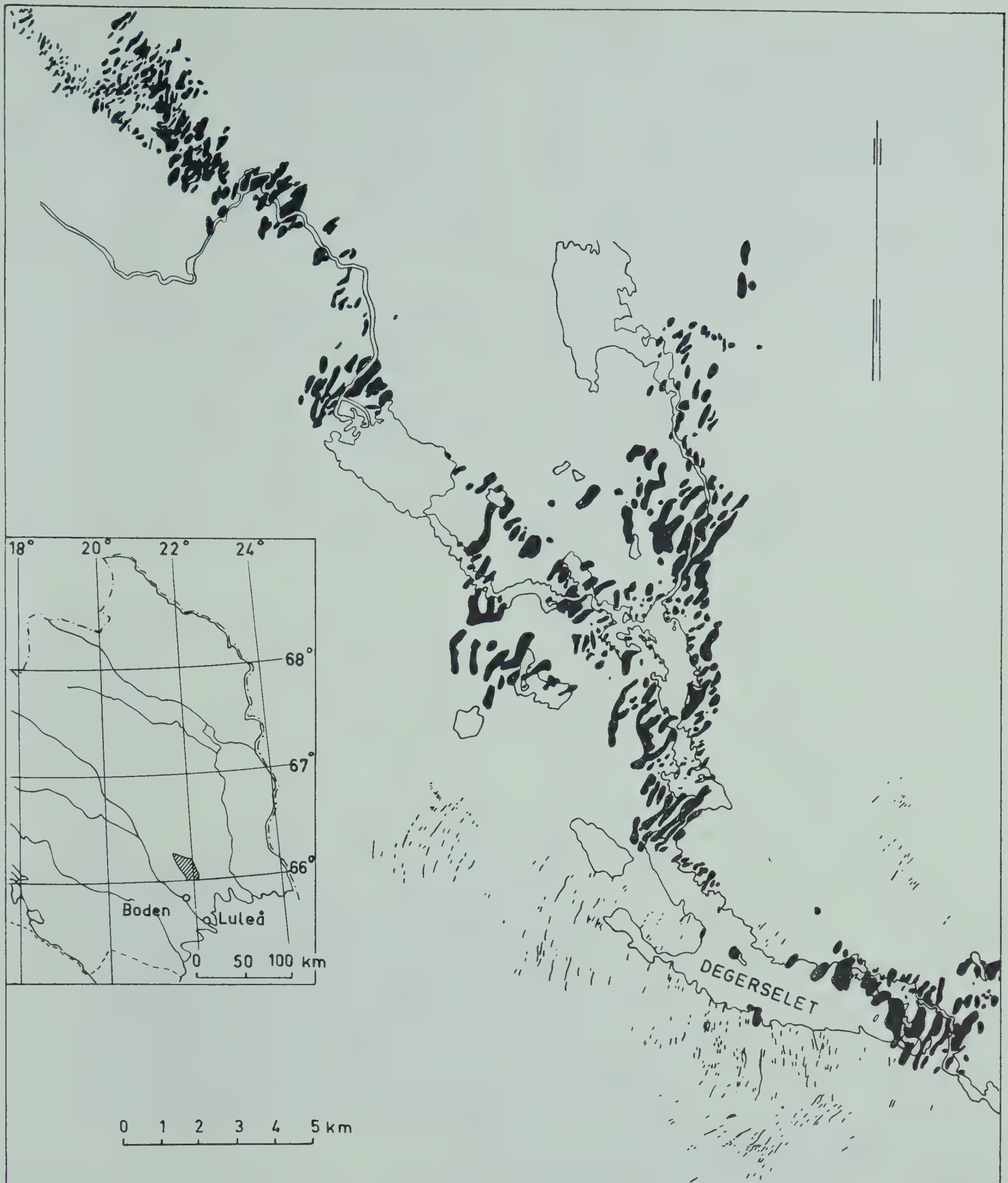


Fig. 8. «Kalixpinnmo» hills (large black areas) and DE GEER moraines (black lines) in a part of the Råne valley (HOPPE 1948).

The most representative areas with "Kalixpinnmo" are the valleys of the Kalix and Råne rivers, but it may also be found in many places in the Lule and Pite valleys, and outside of Norrbotten; even above the highest shoreline. In the Råne valley (fig. 8) the "Kalixpinnmo" hills are concentrated in the lowest part of the valley and change successively on the sides of the valley into typical De Geer moraines through reduction of the sediment content. The "Kalixpinnmo" also replaces the more coarse glacifluvial deposits, eskers, etc.

The "Kalixpinnmo" hills seem to have been formed subglacially, or, to be more correct, submarginally. The normal fast-running meltwater stream was replaced by slowly-flowing water which transported great quantities of the above-mentioned constituent materials, mostly in suspension, but also along the stream bottom. The reason for this replacement is not known exactly, but may have been caused by a combination of the thinning out of the ice, which diminished the hydraulic head; by the pressure of the water outside of the ice-front; and also by the buoyancy of the ice which widened the meltwater channels to broad passages over the lowest parts of the valleys. During the sedimentation process till constituents also were dropped from the roof of ice. The deposits were then packed together, thrust and kneaded by the outermost parts of the ice which were still in motion. According to this explanation the "Kalixpinnmo" hills ought to be described as a type of subaqueous thrust end moraines ("subaquatische Stauchendmoräne", cf. Gripp 1938).—For the formation of the "Kalixpinnmo" it is also important to note that the normal till of this area—i.e. the parent material—has a rather high content of fine sand and coarse silt (cf. curve 7 in fig. 7) and that these grain sizes are the most easily eroded ones (Hjulström 1935, Sundborg 1956).

The general progress of deglaciation in the area below the highest shoreline

By means of De Geer moraines, glacial striae, glacifluvial deposits and "Kalixpinnmo" hills a reconstruction of the positions of the ice terminus at various times can be made. Fig. 9 represents such an attempt, covering a central part of the Norrbotten coastland and the archipelago outside.

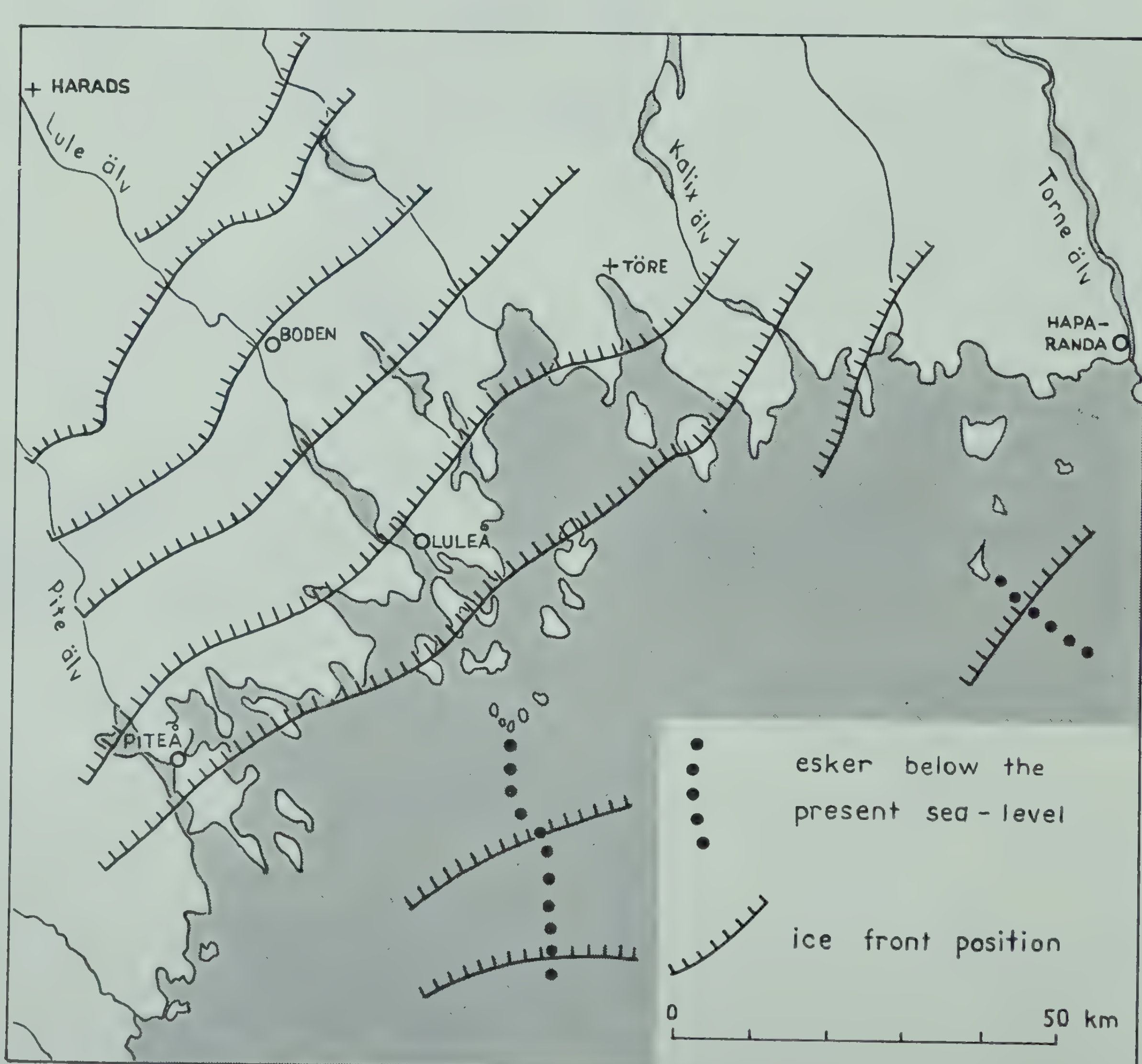


Fig. 9. Deglaciation of southeastern Norrbotten. (From Hoppe 1948, with some complements).

The ice-front ran in a general NE—SW direction, which means that it receded to the northwest, and that the (or one) center of the inland-ice was at that time situated NW of the Norrbotten coastland. During the recession there were many exceptions from this general direction of retreat, caused for instance by changes in the accumulation area of the ice, by the characteristics of the sea depth which might result in "calving bays" or for some other reason. Such deviations are very common in the area just below the highest shore-line, where the topographical influence may have been very strong. (An example is given in Hoppe 1948, p. 40.)

On the basis of pollen and diatom analyses, Fromm (1949, 1960) has dated the deglaciation of the Norrbotten coastland as having taken place approximately 7,000 years B.C. In the development of the Baltic it corresponds to the transition between the Yoldia and Ancylus stages.

Postglacial development

All the land between the highest shoreline and the present sea-level has once been the object of coastal processes. The more exposed parts, especially, have been strongly abraded, so they are now represented by bare rock or residual boulders. Spits and hooks are also common features in such situations. The lower parts of the hills are usually surrounded by stratified, outwards-dipping gravel and sand (= "svallgrus"), which has been swept down from the higher parts. —The effect of the coastal agencies is extremely obvious in connection with the glacifluvial deposits, which have been transformed to a rather high extent.

Parallel to this development there was a strong deposition of sediments, especially in the river valleys. These first formed deep inlets, where finegrained so-called "fjord sediments" accumulated in great thicknesses, while later delta and other river sediments were deposited on top of the older sediments when the mouths of the rivers successively moved outwards.—The total thickness of such postglacial sediments may exceed 50 meters.

GLACIAL LANDFORMS AND DEGLACIATION IN THE AREA ABOVE THE HIGHEST SHORELINE

When trying to follow the progress of the deglaciation in areas above the highest shoreline we must use the evidence offered by other types of landforms and look to other lines of reasoning. Of particular importance in these areas are the channels formed by meltwater. Also significant are the glacial striae, the drumlins, glacifluvial deposits of different kinds, and, where they exist, traces of ice-dammed lakes. Recessional moraines do not appear except in the immediate vicinity of present glaciers but moraines of other types, especially hummocky variants, can help us reconstruct a detailed view of the deglaciation period.

The meltwater channels

Channels eroded by the meltwater from the receding ice occur in all parts of Norrbotten in very great number. Only in the high-mountain areas—where they probably did exist in large number but were later destroyed by mass-movement processes—and in the hummocky-moraine regions, are such channels rarely found. They serve best to illustrate the deglaciation principles when they appear both on both sides and also on the bottom of a valley (Hoppe 1950). In these cases the meltwaters had more or less embraced an ice-tongue, the channels now demonstrating its successive shrinkage and the corresponding recession of the ice-front. One-sided, extremely regular meltwater channels appear in large numbers on some slopes in the mountains (e.g. on Kaisepakte south of Lake Torneträsk; on Larkimvare south of Nikkaluokta; at the Kaitumjaure lakes; and on Ultevis, southeast of Saltoluokta). We assume that the channels in such cases were formed along the margins of the ice and they may indicate the downward gradient of the ice surface (normally varying between a 1 and 3 percent slope). Furthermore the vertical distance between two consecutive channels has been thought of as representing the diminution of the ice during one year (an average value being 3 to 5 meters/year). The last assumption especially is open to criticism. Holdar has shown (1957) that in some areas the vertical distance between two successive channels may reach 25 or 30 meters, which for theoretical reasons can not be considered to represent a single year's shrinkage through melting (Wallén 1957). The interference of lateral moraines may also complicate the picture (observed on Greenland by Schytt (1956) and Hoppe and Schytt in Vestspitsbergen (1957, unpublished).

Some of the meltwater channels are enormous phenomena which may exceed 10 km in length and 50 meters in depth and are often cut deeply into the bedrock. Well-known canyons of this type exist, for instance, in the river basin of the Lule älv, "Lagmansgraven" SE of Jokkmokk, Ahos kårso (kårso: Lappish word meaning 'canyon') at Saltoluokta, and a series of channels entering the valley of the Stora Lule älv from the northeast downstream of Porjus. The deepest of the latter canyons, Muddus kårso, is now also used by a river; but it opens on an undissected terrace at the highest shoreline, thus demonstrating that the postglacial erosion has been insignificant. How such gigantic canyons could have been formed during such relatively short periods of time—in most cases probably only tens or hundreds of years—is an open question. Probably the meltwater streams ran subglacially under considerable hydrostatic pressure at a very high speed, perhaps high enough to cause cavitation. In many cases it has been possible to show that they followed zones of weakness of the bedrock (e.g. Lidnokursu between Gällivare and Kiruna, Hoppe 1950). As is the case with the well-known Bardu (or Sördals) Canyon NW of Lake Torneträsk (Holdar 1952), the possibility that a canyon has been used during more than one glaciation and therefore subjected to ice erosion as well can not be excluded.

Glacial striae

On the whole, well-preserved glacial striation will not often be observed on bedrock outcrops. In the wide Sarek area, for instance, only some twenty striae localities have been observed; in other regions there are hundreds of square kilometers without any observation at all. On the other hand there are areas—especially with fine-grained (and quite soft) rocks—where excellent examples of scratching can be observed. In these areas it is not uncommon to find two or more sets of striae running in different directions on the same surface. Great difficulties often arise in the relative dating of such cases.

When they can be correlated with nearby meltwater channels, the youngest striae in a locality are often seen to correspond to a late stage of the deglaciation when the ice over the area formed only a rather thin and narrow tongue. This means that the ice was active practically to the time of its disappearance and that there were no separations of bodies of dead ice (with probable exceptions in certain topographical situations).

Moraines

Drumlins

Drumlins are common features especially in the area between the highest shoreline and the high mountains. Most of them form elongated and quite irregular ridges, seldom more than ten meters high and with a length of up to 500 meters or even more. They usually consist of a normal, rather sandy till. When the stone content is not too high, i.e. when the stones have a quite free position, they show a preferred orientation in the direction of the ridge. This must be the result of subglacial deposition which either took place directly from the ice through lodgment, or by the forcing of water-soaked moraine into elongated cavities; the two possibilities which would seem to most adequately apply to the drumlins in Norrbotten. Also stones situated just under the surface show this preferred orientation, which suggests that the ice must have been practically free of superglacial drift in the drumlin areas. Wright has also discovered (1957) a preferred up-glacier plunge of the stones in a drumlin field in Minnesota, which makes it possible to decide in which direction (of two alternatives) the ice moved. No such preferred plunge has yet been observed in the Norrbotten drumlins, nor have they an apparent regular asymmetry which would permit the application of this approach in the determination of direction; this is, however, important only in a few cases as direction can ordinarily be determined through other methods.—In some of the drumlins cores of bedrock have been noticed; more common, however, is that they lack such cores.

As far as has been observed, the drumlins are always oriented in the direction of the *last* ice movement as demonstrated by the youngest striae (Hoppe 1951, 1957). This means that they were formed during a very late stage of the glaciation at a time when the receding ice front was perhaps only tens of kilometers away.

In some areas there is a striation of till surfaces, perhaps better described as fluting than drumlinization. Owing to rather small variations in height and a very straight and regular lineation, the fluting is often difficult to observe from the ground but is easily spotted on aerial photographs.

A similar fluting is a normal feature in front of existing retreating glaciers in northern Sweden, as well as elsewere in the world (cf. Hoppe-Schytt 1953).

Hummocky moraines

In many areas in northern Sweden, especially where the till cover is very thick, the landscape has a hummocky appearance. Besides those areas with no specific characteristics, two well-defined categories of hummocky moraines have so far been described. One of them, called the *Rogen moraine* after the location of a well-known example (Lundqvist 1937, Mannerfelt 1945, Hoppe 1952), appears generally in mountain valleys, although it occasionally also can extend over more non-dissected landscapes. It is distinguished by ridges up to 30 to 40 meters in height—which would suggest that it is not quite correct to describe it as a "hummocky" moraine—most of the ridges running transverse to the valley direction. It typically has a very high content of big boulders and a rather low percentage of fine grain sizes.

The Rogen moraine is still unexplained. It is important to note that in some cases eskers appear superimposed on the ridges, and also that ice-margin meltwater channels may occur on higher parts of the ridges; both facts speaking for a subglacial formation. In a few valleys an asymmetrical cross-profile of the ridges has been observed with a steeper distal (down-glacier) side.

The *Veiki moraine* (Hoppe 1952, 1957) consists typically of three elements: moraine plateaus, rim ridges on the plateaus and hollows, often with ponds and peat-bogs, between them. A transverse stone orientation is very typical in the rim ridges, while the stones of the plateaus, as far as has been studied, seem to be oriented quite irregularly. Often small ridges or terraces—having the same stone orientation as the rim ridges—appear on the slopes down to the hollows. In other cases isolated ridges occur without any combination with plateaus; such ridges can consist of a whole complex of smaller ridges, one above the other, all with the characteristic stone orientation. The rim ridges, as well as other categories of ridges, are very similar to the De Geer moraines, both as to aspect and texture.

In many regions the Veiki moraine alternates with drumlin areas, which suggests that also the former was formed subglacially. This hypothesis has been conclusively proved by this author in his paper of 1957 which describes a locality in which the drumlins are superimposed on moraine plateaus. For this locality the following manner of formation is presumed: (1) an uneven deposition of moraine, whereby plateaus and hollows were formed, (2) a formation of drumlins on the surface of the plateaus, (3) a squeezing of water-soaked moraine from the hollows against the plateaus (in this manner the rim ridges were built up). The ice was still moving during (1) and (2), dynamically dead during (3). The intimate connection of the genesis of the plateaus, drumlins and rim ridges is manifested by their practically identical grain size curves (Hoppe 1959, p. 312).

The Veiki moraine has been found up to now only in the northern parts of Sweden. C. P. Gravenor, A. Stalker, and other Canadian geologists have shown the author similar moraine landscapes with plateaus and rim ridges situated in Alberta. The surface of the plateaus there, however, is quite often covered with varved sediments, and a core of folded and even overturned bedrock in the plateaus and ridges is a normal feature. It is therefore doubtful that the hummocky moraine of Alberta has anything in common genetically with the Veiki moraine.

Glacifluvial deposits

Especially outside of the mountain regions, glacifluvial deposits form more or less continuous chains; in the mountains they appear more sporadically. Eskers and associated features dominate; in many places outwash sediments, formerly often overlooked, are also common. In the more mountainous regions glacifluvial deltas and terraces are widespread.

Eskers

Many of the eskers which exist below the highest shoreline can be followed from this region, once subaqueous, into the supraaqeous area. There they often continue in sections of 5 or 10 kilometers or more, but between these there are breaches which often coincide with higher terrain features, where the eskers frequently are replaced by erosional zones.—The bigger eskers usually have a rather flat and broad crest, which we suppose had formed the bottom of the meltwater streams, successively raised during the deposition. The author agrees with Tanner (1938) that the

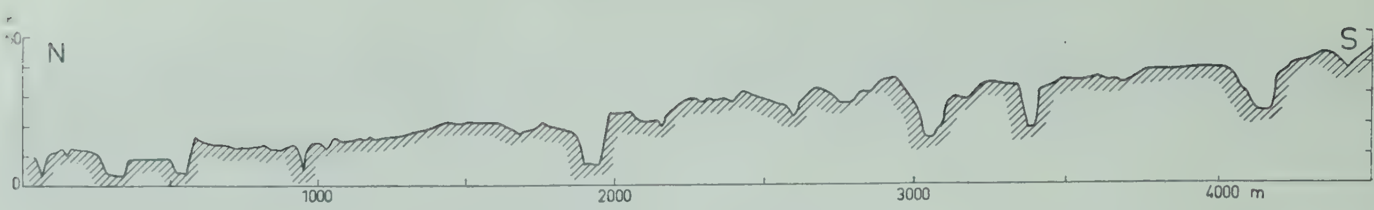


Fig. 10. Profile along the crest of the esker Hammasharju. (Surveyed by J. A. NILSSON.)

so-called "goat-ridges" with very sharp and narrow crests—normally occuring where the eskers are not as high as the average level—represent a secondary shape produced by the sliding away of the sides.—The supraaqueous eskers of Norrbotten are built up mainly of stratified sand, gravel and stone. Occasionally heaps of boulders occur, in other cases very fine materials such as silt, which must have been transported in suspension, have been observed both in the cores of eskers and as thin covers on top of them.

In the opinion of Tanner the supraaqueous eskers of northern Fennoscandia were formed in superglacial stream valleys. His basic argument was a theory that a thick superglacial moraine cover supplied the material for the eskers; this theory, however, can hardly be correct—as we have already indicated in the discussion of the drumlins and the "Veiki moraine". Also the normally strong influence of the topography and the well-preserved stratification in the eskers argue against a superglacial—and also an englacial—formation and for a subglacial deposition. Some significant insights into this question have been obtained through the study of a small esker called Hammasharju situated southeast of Gällivare. It has a series of gaps (fig. 10), most or all of which apparently were made by running water. The case seems to be very similar to an esker in Denmark described by Andersen (quoted in Flint 1957, p. 158). Hammasharju is situated on terrain which slopes in a proximal direction (= up-glacier); the crest of the esker sloping in the same direction. Formed in a subglacial tunnel, the esker was subsequently exposed. A superglacial stream was then superimposed across the esker, its course moving gradually in a proximal direction as the ice waned, thereby cutting the gaps, one after

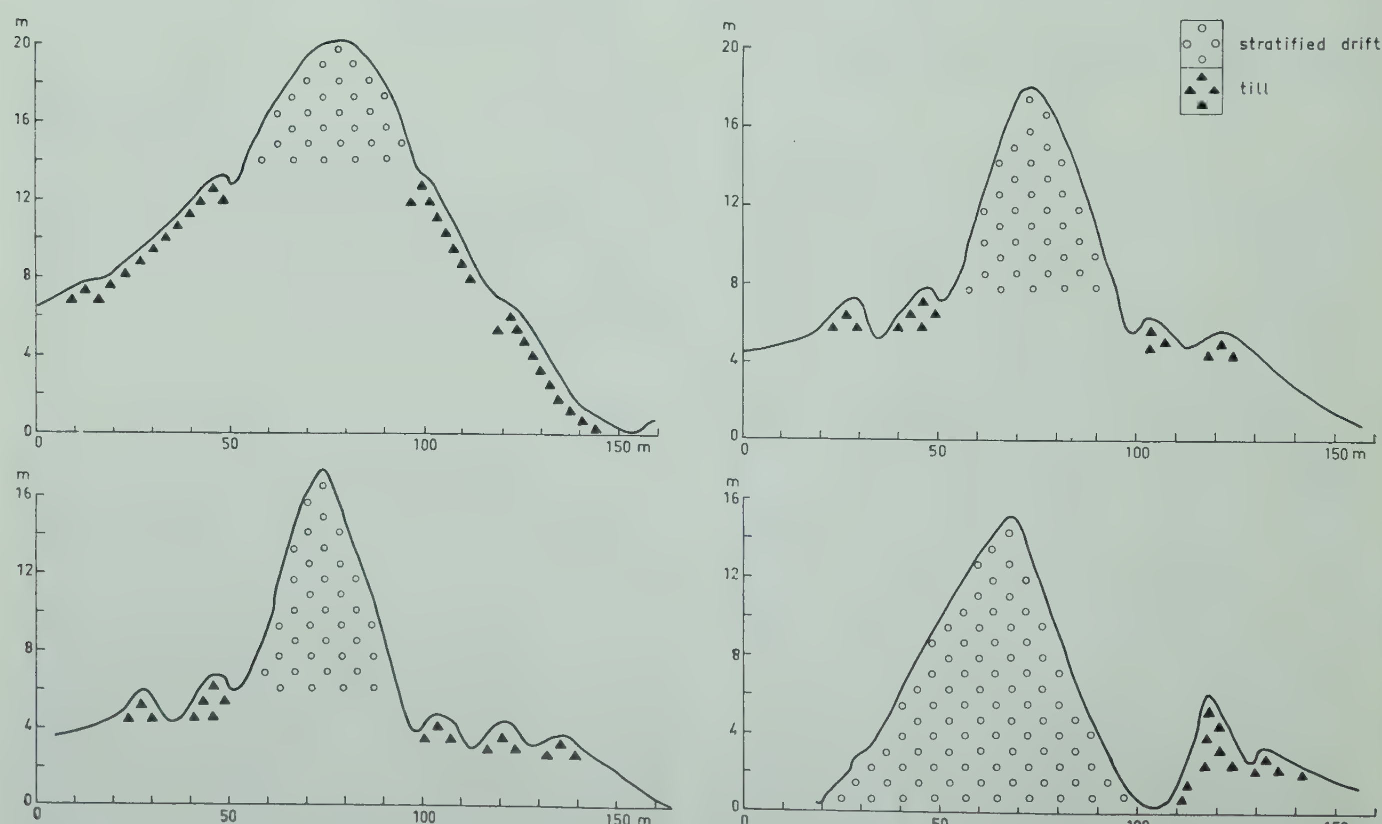


Fig. 11. Cross profiles of Hammasharju. (Surveyed by G. HOPPE and J. A. NILSSON).

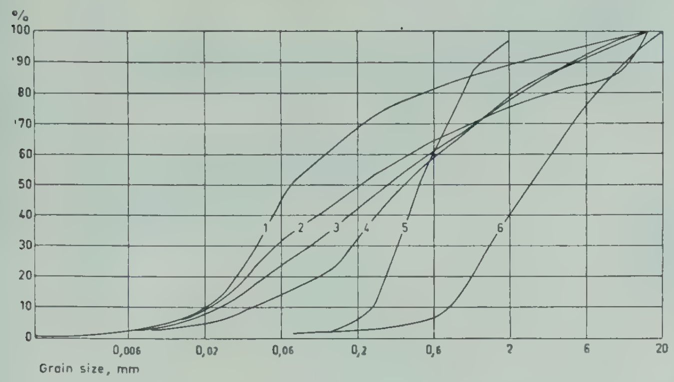


Fig. 12. Grain size distribution of ridged till (1—4) and stratified drift (5—6) of Hammasharju.

another. Hammasharju, however, is remarkable in another respect. In many places it is accompanied on one or both sides by small parallel till ridges (fig. 11, grain size curves in fig. 12), similar to the rim- and other ridges of the Veiki moraine and also to the De Geer moraines. These similarities also include a preferably transverse orientation of the elongated stones. If we apply the same interpretation that we developed for the seemingly related features above, the till ridges are shown to have been formed by the forcing of watersoaked moraine from below the ice against the tunnel, which probably had widened a little through melting after the deposition of the stratified drift. The esker with its parallel moraines then offers a variant of what the Germans call "Aufpressungsoser" (first described by Korn 1910), the difference being that in the latter, the till has been squeezed in under the esker, lifting it and forming a central core within the esker.

Outwash plains

Taking advantage of advances in our understanding of the different kinds of glacifluvial deposits, especially through the study of aerial photographs, taken over practically the whole of Norrbotten, it has been possible during the last few years to identify quite a large number of plain areas as consisting of outwash sediments. Only one of these outwash plains, called the "Petsaure sandur", has up till now been investigated (Hoppe *et aliis* 1959). Situated in a narrow mountain valley, this plain is about 9 km long and in places up to one km. wide, thus belonging to the kind of outwash plains characterized as valley trains (Flint 1957, p. 139). It has the typical pattern of braided streams. In its headward part the gradient exceeds 1 %, diminishing to values of less than 0.5 %, in the lowest part. There are signs that the sediments have been built up at successive positions of the front as one would expect since it was formed while the ice was receding. A number of kettles, both large and small, pit the plain: most of these probably reflect the irregular terminal zone of the glacier, which was buried in the drift. The up-valley part of the plain consists to a large extent of boulders and stones, while the lowest part, which probably was deposited in Lake Petsaure when it was at a higher water-level than now, is sandy. Bimodal grain size distributions with a pronounced minimum in the 2 to 8 mm range are typical for a large part of the "Petsaure sandur".

Deltas and terraces

Where meltwater streams from receding ice masses in higher areas came down into lower-lying valleys still occupied by glacier tongues, a damming up of lateral lakes often occurred and water-transported drift was deposited in them. This glacifluvial material now forms terraces and deltas on the valley sides. The surface of the glacier tongue was continually shrinking during the deposition of the drift, and there is now often a whole series of deltas and terraces, one below the other. Many of the deltas are pitted, proving that the water was dammed by ice in the immediate proximity. Because of a short transport distance the sorting coefficient of the drift usually is low.

Well-known ice-dammed lake terraces and deltas appear in many places around Lake Torneträsk, for example, in Pässisvagge, and in the Stora Lule älv valley, at Saltoluokta and northwest of the Akka mountains. This phenomenon is especially common where hanging valleys run into main valleys; a large number of examples are found in the Sarek mountains. A very important chain of terraces is found alongside the Könkämä river, the river which forms the frontier with Finland along Sweden's northern tip.

THE GENERAL PROGRESS OF DEGLACIATION IN THE AREA ABOVE THE HIGHEST SHORELINE

Until the last ten years it was generally agreed that during the latter stages of deglaciation in northern Sweden the last ice remnants were

Fig. 13. Main deglaciation landforms in the area between Torneträsk and Stora Lulevatten.

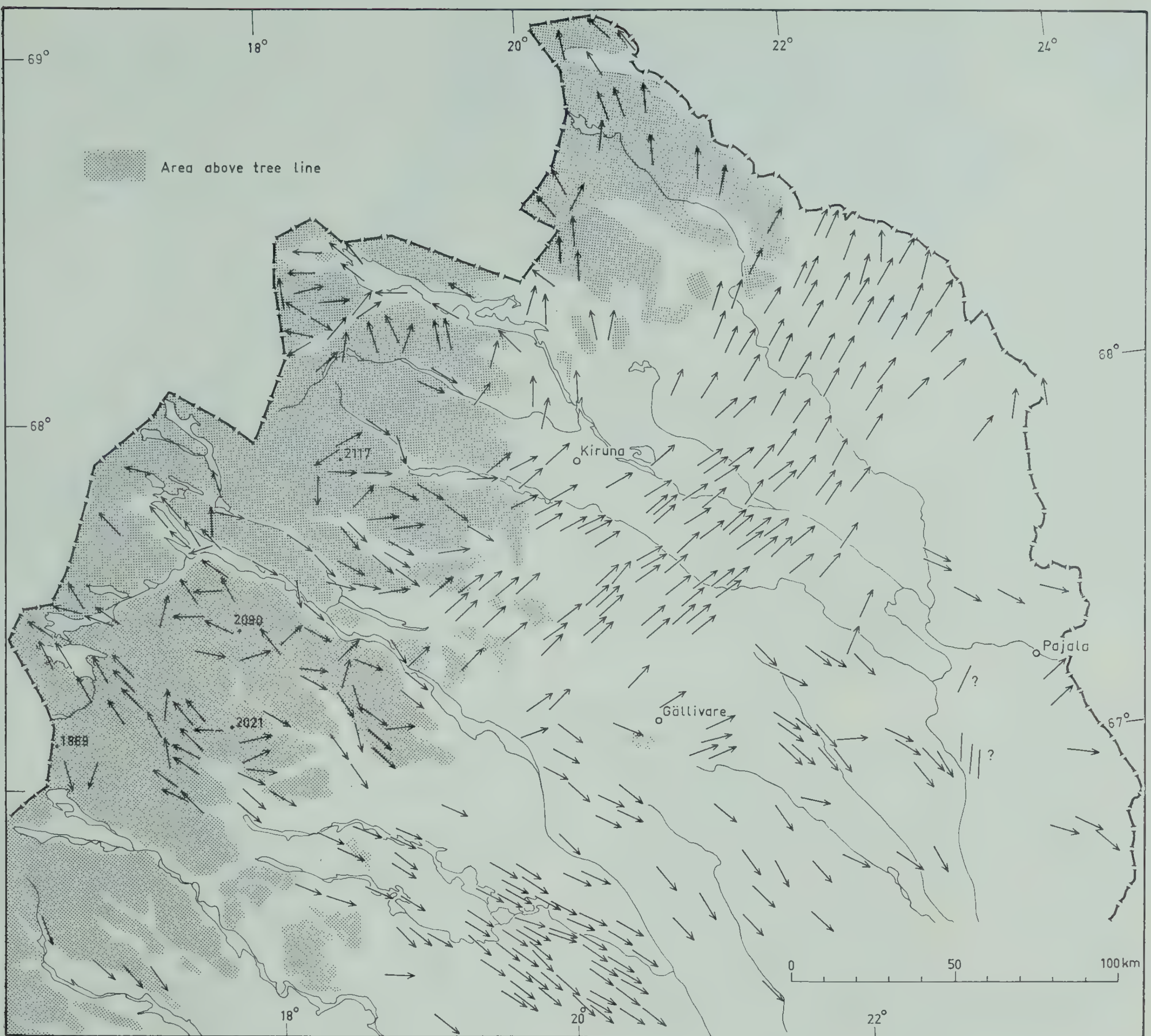


Fig. 14. Last ice movement directions in northern Norrbotten. The map is mainly based on the author's studies of glacial striae, drumlins, glacifluvial deposits and channels. Gaps are caused by the lack of such features and/or field studies. At a couple of places it has not been possible to decide whether the ice has moved from one or the opposite direction.

situated east of the high mountains and dammed up a whole series of glacial lakes. As evidence to support this theory, attention was drawn to a large number of strandlines as well as glacifluvial deltas and terraces in the mountain regions. As the hummocky moraines (''dead-ice-moraines") were thought to have resulted from the melting of the ice masses without any movement at all, the theory about the situation of the last ice remnants was coupled with the supposition that these were practically dead.

Investigations during the past 10 or 15 years more comprehensive both as to the number of landform categories studied, and especially as to the area, have discredited the above concept (Hoppe in Hoppe-Liljequist 1956, Holdar 1957). Aerial photography has made possible a quite complete mapping of, for instance, meltwater channels, eskers and other glacifluvial deposits, drumlins and other kinds of moraines, etc. Some of the results of these recent investigations are shown—in small scale and therefore necessarily rather schematically—on the map in fig. 13. This map covers a strip between Lake Torneträsk in the north and the valley of the Stora Lule älv in the south and includes both high mountains—the Kebnekajse complex—and forested lands east of the mountain chain. This area has been of

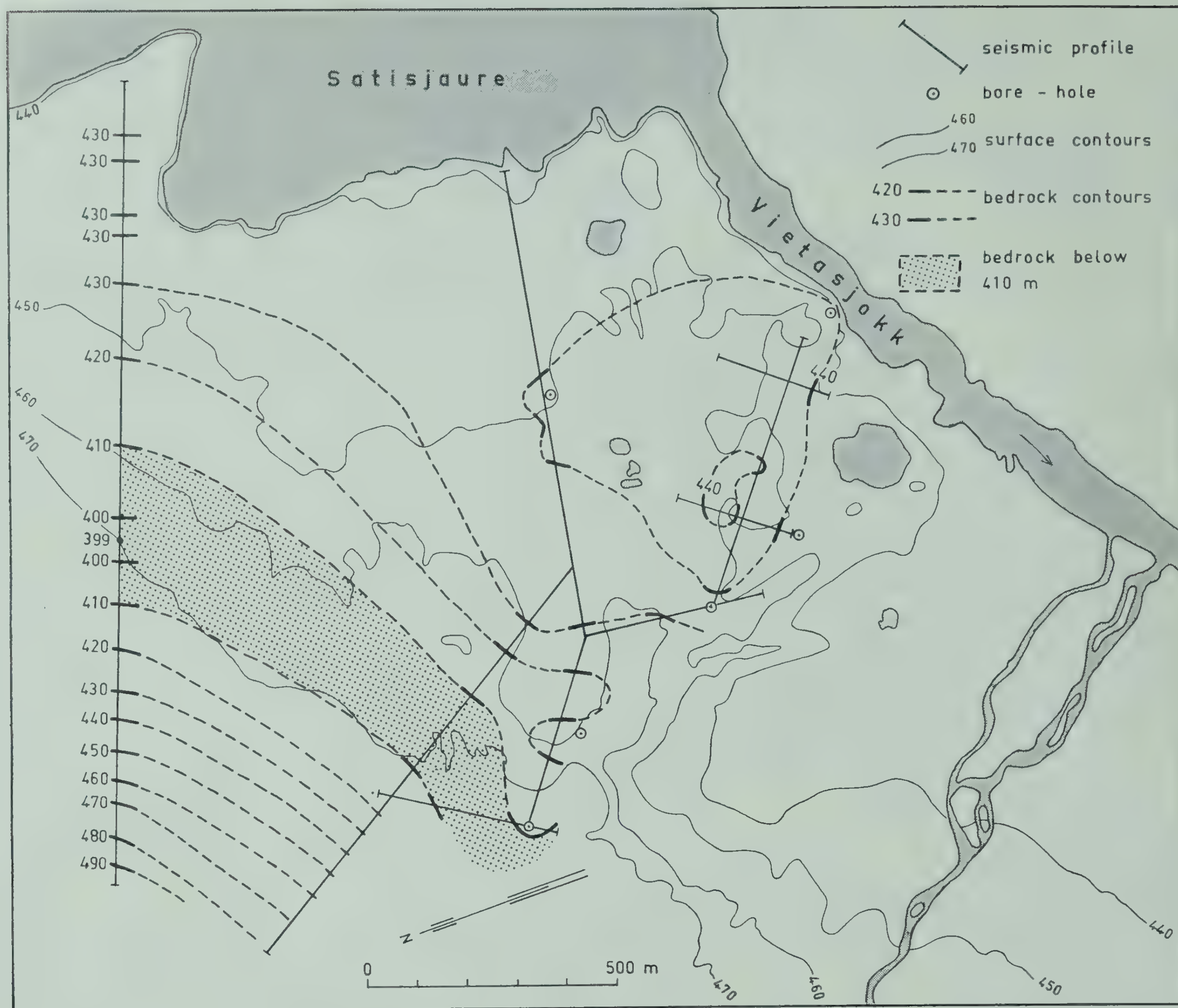


Fig. 15. Subsurface bedrock channel close to Satisjaure, covered with 50—70 m of glacial deposits of various origin.

notable importance in respect to the conclusions made below. These conclusions are shown in the map, fig. 14, demonstrating the most recent ice movement directions in northern Sweden.[1]

At the maximum extent of the last glaciation the center of the ice sheet was situated far to the east, probably over the Gulf of Bothnia; towards the final stage the center moved westwards to a position over the mountains. The reason seems to have been a return of the meteorological situation to conditions similar to those at present (and probably also similar to those of interglacial times); the cyclonic tracks once more began to cross northern Scandinavia in large numbers, bringing much precipitation (snow) over the western side with the westerly winds (Hoppe-Liljequist 1956).—When the ice front was still in eastern Norrbotten, the center of the ice sheet seems to have been situated over the uppermost part of the river basin of the Lule älv; perhaps it can be localized to the Sarek area, where moreover the precipitation at the present time seems to be the highest in Sweden. During the continuous retreat of the ice border and the corresponding sinking of the surface of the whole ice sheet for climatological and dynamic reasons, this center was replaced by a number of secondary centers, each forming a focus of glacial activity. Such centers have been identified for instance in

[1] It must be emphasized that the movements indicated are metachronous in much the same manner as one represents glacial striae on maps covering large areas. This map is based mainly on the studies of the author, supplemented for some areas, however, by material collected by other scientists. There are still some voids, on account of the absence of aerial photographs, detailed field studies etc.

the Sarek, Sulitelma, Kebnekajse and Kårsa mountain areas. Each of these centers probably formed a glacier cap which was transformed first to a transection glacier then further to valley and cirque glaciers, most of them situated in the localities of present glaciers. During this development the effects of local topography came to exert a stronger and stronger influence on the directions of the ice movements. In the mountain areas there were often changes in diametrically opposite directions as a result of the changes and divisions of the different centers. Examples of such a development are given in the papers of Holdar 1957 and Hoppe *et aliis* 1959.—Referring to the curve showing the rate of present land upheaval (fig. 4) it would seem to show a secondary maximum in the mountain area (cf. Scherman 1959). If this is correct, it can perhaps be correlated with the last ice center over the mountains.

The thickness of the cover of deposits of glacial origin varies greatly. As a rule it is very thin or totally lacking on the higher parts of hills and mountains. Similarly in the areas alongside the main water divide of northern Scandinavia deposits are sparse or lacking even in low terrain. On the other hand the thickness may be considerable east of the water divide, for instance in the hummocky moraine areas and in many valleys. The soil cover may often conceal interesting morphological details of the bedrock surface. An example is given in fig. 15, showing an area about 15 km north of Saltoluokta. At this site the State Power Board has made a series of seismic profiles and borings from which the main features of the bedrock surface can be reconstructed. Under a 50 to 70 meter cover of loose deposits a valley has been discovered. Its continuation, age—preglacial or interglacial—and the forces which formed it are still unknown.

References

Beskow, G., 1935. Praktiska och kvartärgeologiska resultat av grusinventeringen i Norrbottens län. Geol. För. Förh., 57.

De Geer, G., 1889. Ändmoräner i trakten mellan Spånga och Sundbyberg. Geol. För. Förh., 11.

— 1940. Geochronologia Suecica Principles. K. Svenska Vetenskapsakademiens handl., 3 Serien, 18: 6.

Enequist, G., 1944. Dynstudier i Lule skärgård. Geographica, 15.

— 1946. Isälvsavlagringar i Luledalen nedanför Hednoret. Geogr. Annaler, 28.

Field, Jr, W. O., 1947. Glacier recession in Muir Inlet, Glacier Bay, Alaska. Geogr. Review, 37.

Flint, R. F., 1957. Glacial and Pleistocene geology.

Fromm, E., 1949. Datering av den senglaciala utvecklingen i Norrbottens kustland. Geol. För. Förh., 71.

— 1960. An interglacial peat at Ale near Luleå, Northern Sweden. Sver. Geol. Unders., Ser. C, 574.

Gripp, K., 1938. Endmoränen. Intern. Geogr. Congr., Amsterdam 1938. Compt. rendus, 2: 2a.

Hast, N., 1959. Mätning av bergtryck i gruvor. Ingeniörsvetenskapsakademien, FKO-medd., 30.

Hjulström, F., 1935. Studies of the morphological activity of rivers as illustrated by the river Fyris. Bull. Geol. Inst. Upsala, 25.

Holdar, C. G., 1952. Problemet Torne-issjön. Geogr. Annaler, 34.

— 1957. Deglaciationsförloppet i Torneträsk-området efter senaste nedisningsperioden, med vissa tillbakablickar och regionala jämförelser. Geol. För. Förh., 79.

Hoppe, G., 1948. Isrecessionen från Norrbottens kustland i belysning av de glaciala formelementen. Geographica, 20.

— 1950. Några exempel på glacifluvial dränering från det inre Norrbotten. Geogr. Annaler, 32.

Hoppe, G., 1951. Drumlins i nordöstra Norrbotten. Geogr. Annaler, 33.

— 1952. Hummocky moraine regions with special reference to the interior of Norrbotten. Geogr. Annaler, 34.

— 1957. Problems of glacial morphology and the Ice Age. Geogr. Annaler, 39.

— 1959. Om flygbildstolkning vid jordartskartering jämte några erfarenheter från jordartskartering i norra Lappland. Geol. För. Förh., 81.

— , Kindblom, B.-O., Klein, K., Klingström, A. and E., 1959. Glacialmorfologi och isrörelser i ett lappländskt fjällområde. Geogr. Annaler, 41.

— and Liljequist, G. H., 1956. Det sista nedisningsförloppet i Nordeuropa och dess meteorologiska bakgrund. Ymer. 76.

— and Schytt, V., 1953. Some observations on fluted moraine surfaces. Geogr. Annaler, 35.

Järnefors, B., 1956. Isrecessionen inom Uppsalaområdet. Geol. För. Förh., 78.

Korn, J., 1910. Erläuterungen zur Geologischen Karte von Preussen und benachbarten Bundesstaaten. Lief. 149, Blatt Marienfliess.

Ljungner, E., 1949. East-west balance of the Quaternary ice caps in Patagonia and Scandinavia. Bull. Geol. Inst. Upsala, 33.

Lundqvist, G., 1937. Sjösediment från Rogenområdet i Härjedalen. Sver. Geol. Unders., Ser. C, 408.

— 1948. Blockens orientering i olika jordarter. Sver. Geol. Unders., Ser. C, 497.

Mannerfelt, C. M., 1945. Några glacialmorfologiska formelement och deras vittnesbörd om inlandsisens avsmältningsmekanik i svensk och norsk fjällterräng. Geogr. Annaler, 27.

Mawdsley, J. B., 1936. The wash-board moraines of the Opawica-Chibougamau area, Quebec. Royal Soc. Canada Ser. 3, 30: 4.

b. Since the verticals have not been rectified and since the existing maps are of a varying quality, both the actual position and the shape of the glaciers may not be quite true as portrayed on the map. This inaccuracy should, however, not be too serious at the scale used here, 1:500,000.

c. All area measurements have been made on un-rectified photographs at a scale of *about* 1:65,000 or *about* 1:35,000. Through comparisons with existing maps a correction has always been applied to the approximate scale, but it has very seldom been possible to adjust for differences in altitude within the picture. We do not believe, however, that the relative error is great as long as the glacier-covered area of a whole massif is considered.

The map (Fig. 1) shows the 218 glaciers found between 67°N and 68°.4 N on the following topographical sheets: 3 Abisko, 7 Akka, 8 Kebnekajse, 12 Sulitelma and 13 Stora Sjöfallet. Besides these (see inset on Fig. 1) there are three in the northernmost corner of the country and another two northeast of Riksgränsen, one is found on Staika (66°59′ N, 17°05′ E), two in the Arjeplog mountains, one just west of Ikesjaure (66°51′N, 16°00′E), five in the massif Norra Storfjället (65°55′N, 15°15′E), one on the nearby Ammarfjället, three in Sylarna (63°01′N, 12°05′E) and one on Helagsfjället (62°55′N, 12°30′E). *According to the present inventory Sweden thus has 237 glaciers covering a total area*[1] *of 310 km²*.

Regional distribution

As indicated in the previous paragraph, 220 out of the 237 identified glaciers are found in the mountains north of the 67th parallel and south of 68°35′N.

Their distribution according to topographic sheets is as follows:

Sheet	Number of glaciers	Total glacier area in km²
3 Abisko	52	42
7 Akka	16	10
8 Kebnekajse	53	48
12 Sulitelma	38	82
13 St. Sjöfallet	61	121
Total	220	303

[1] Only the Swedish parts of glaciers on the Norwegian border have been included in the total area.

Certain mountain areas stand out as important glacier centres. In Sulitelma, i.e. the area along the border between the lakes Virihaure and Peskehaure, we have 2 smaller and 3 large glaciers: Stuorrajekna, Salajekna and Ålmallojekna. According to Westman's photogrammetric map of 1889, Stuorrajekna covered 14.7 km², the Swedish part of Salajekna was 15.9 km² and Ålmallojekna 22 km². The present study gives for Stuorrajekna 12.0 km², Salajekna 13.4 km² and Ålmallojekna 10.8 km² (12.9 km² with the Norwegian part included). The glacier-covered area amounts to 38 km².

The most glacierized area of Sweden is found in the Sarek national park. If the Sarek Mountains in a wider sense be defined as the area from 67°08′N (south side of Pårtetjåkko) to 67°36′N (northern slopes of Akka) and from 17°15′E to 18°05′E this high mountain district includes 100 glaciers with a total area of 171 km². Judged from the air photographs the largest glaciers of Sweden can be found on either side of the valley Sarvesvagge (67°15′N). On the south side the Pårte glacier seems to hold the record for the whole country with its 14.1 km² and on the north side we find the Jokotjkaska glacier, 13.7 km², and the South Ålkatj glacier, 8.3 km². Our figure for the Mikka glacier, the best known glacier in Sarek, is 8.0 km².

The Kebnekajse mountains at about 67°55′N, 18°30′E offer another large concentration of glaciers. The "inner Kebnekajse", delimited by the valleys Ladtjovagge, Vistasvagge, Kaskasavagge and Tjäktjavagge, has 17 glaciers covering an area of 18.6 km², while the Kebnekajse Mountains as they are often defined by the triangle between Ladtjovagge, Vistasvagge and Tjäktjavagge (continued in Allesvagge) contains 44 glaciers with a total area of 41.3 km².

Further to the north large glaciers are found in Mårmantjåkko, where the largest one is 4.3 km², around Kåtotjåkko (largest 2.9 km²) and the northernmost glacier of any size is the Kårsa glacier at the head of Kårsavagge, near the lake Torneträsk.

Size distribution

Several of the glaciers are very small and some would definitely be classified as patches of dead ice if a detailed field reconnaissance could have been carried out. This is, however, not so serious since these patches, especially if delimited

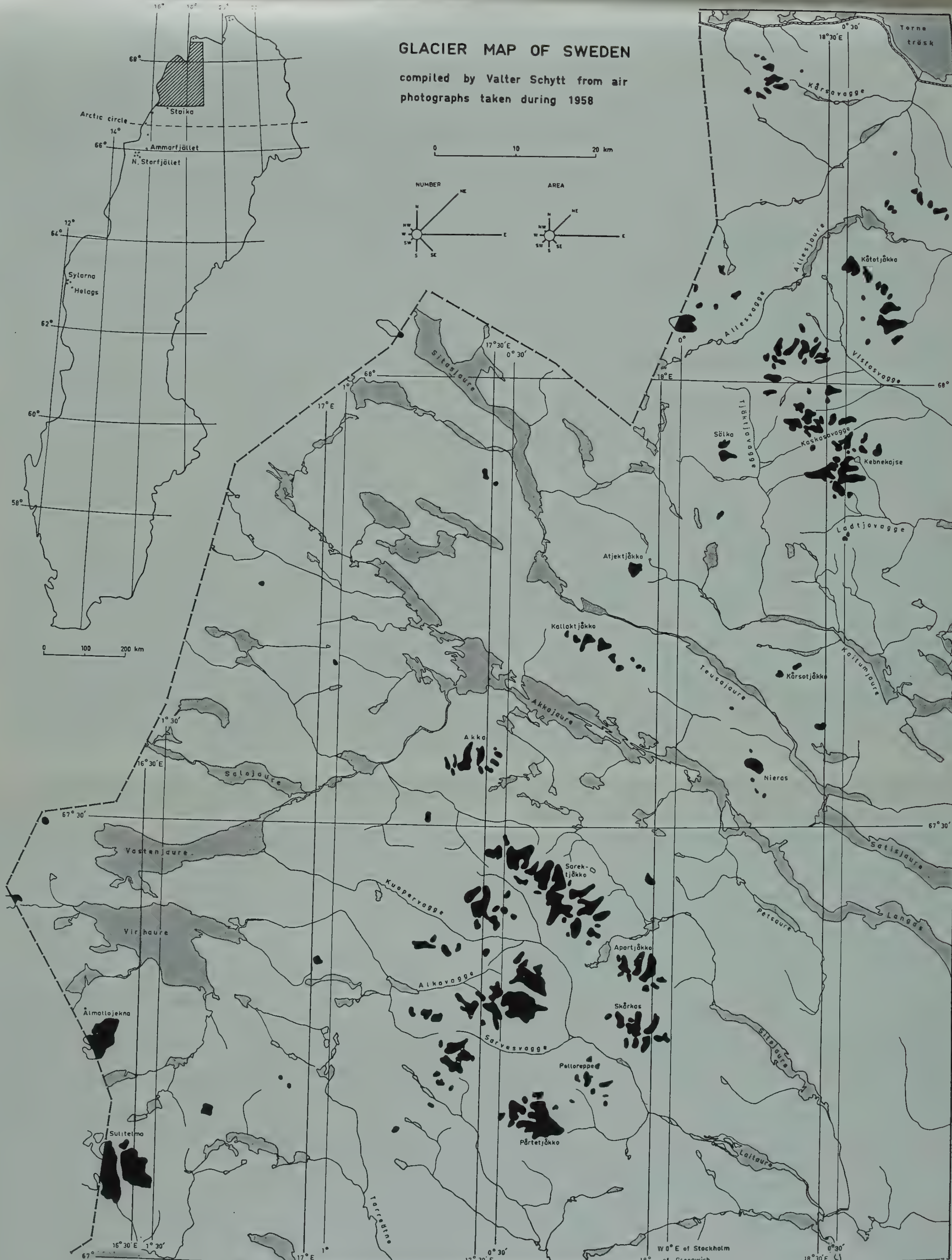
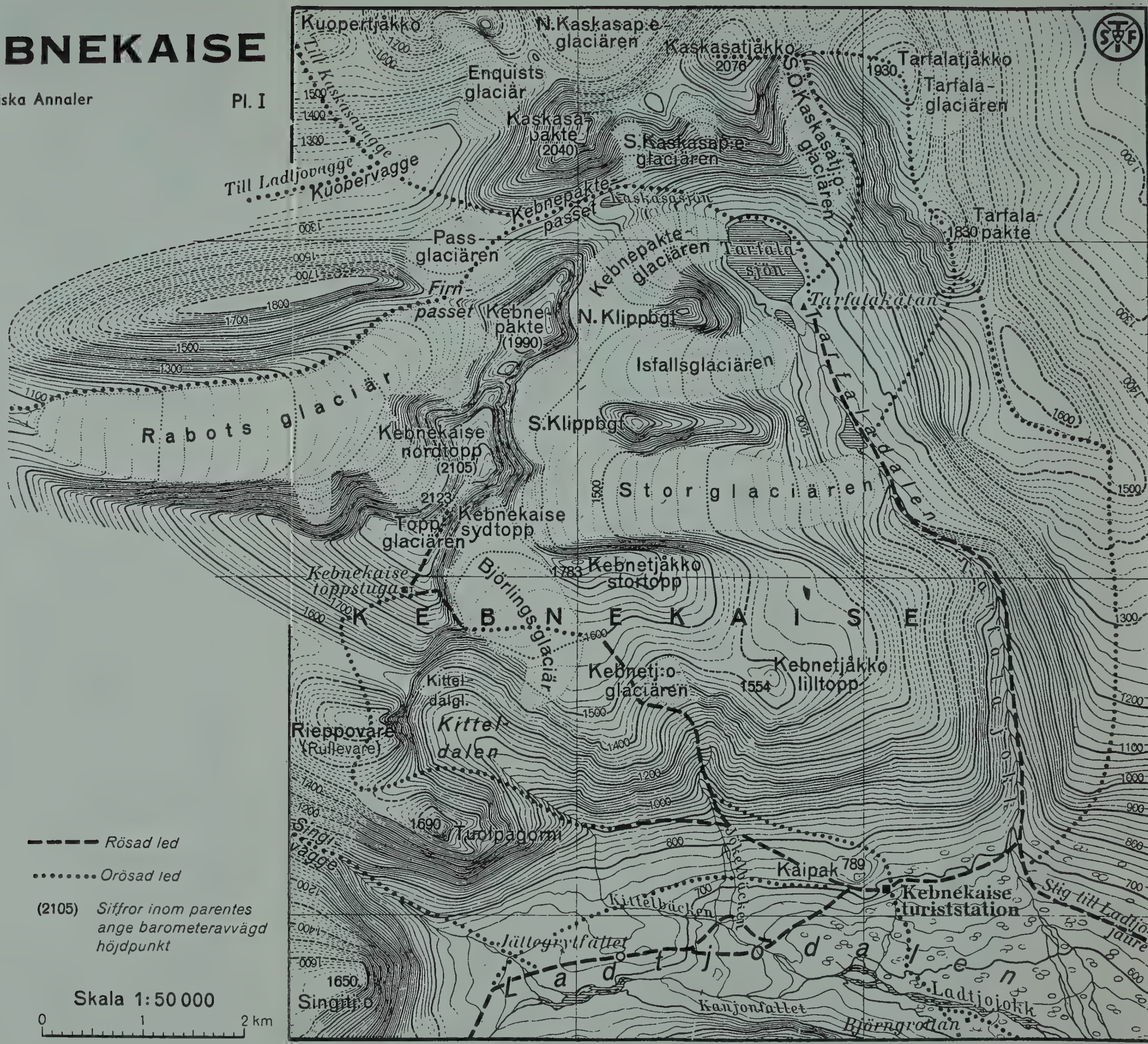


Fig. 1. Kartan för spridning godkänd i Rikets allmänna kartverk den 14 juni 1960.

KEBNEKAISE

Geografiska Annaler Pl. I



Publicerad med tillstånd av Svenska Turistföreningen.
För publicering godkänd i Rikets allmänna kartverk den 8 december 1951.

A.-B. KARTOGRAFISKA INSTITUTET
ESSELTE AB STOCKHOLM 1951

by terminal moraines, represent real glaciers, even though dead at the present time. It is more serious if large perennial snow-fields have been mistaken for glaciers. The probability of such mistakes decreases with increasing area and should have its maximum in the category 0—0.5 km².

The size distribution of all Swedish glaciers is as follows:

Area in km²	<0.15	0.15—0.50	0.5—1.0	1.0—2.0	2.0—4.0	4.0—10.0	>10.0
Number of glaciers	14	92	58	34	23	11	5
Percentage	6	39	24	14	10	5	2

45 per cent of the number of glaciers are thus less than 0.5 km² (19 per cent are 0.2 km² or less) and an absolute error of ± 10 in the total number of glaciers[1] is probably a fair estimate. The area of all glaciers smaller than 0.5 km² does not amount to more than 28 km², i.e. 9 per cent of the total ice-covered area. On the other hand, the 39 glaciers larger than 2.0 km² cover 196 km², i.e. 17 per cent of the number of glaciers correspond to 63 per cent of the glacierized area.

Orientation

It is a known fact that a great majority of Swedish glaciers are situated on the eastern and northeastern slopes of the mountains, i.e. to the lee of the prevailing storm winds (Enquist, 1916).

During the present inventory notes were made as to the orientation of each glacier plotted on the map. Even here it was sometimes difficult to be consistent. Different parts of a glacier may be oriented in different directions. For example, several glaciers move to the north or to the south from an accumulation area due east of a mountain ridge. In the present study we have paid most attention to the orientation of the accumulation area. An easterly firn-field drained to northeast has been listed as an E-glacier, while it has been called a NE-glacier when the tongue has moved due north[2].

For the whole country we have obtained the following distribution:

The two dominating factors influencing the orientation are thus prevailing storm winds and exposure to solar radiation. The winds favour E and NE[3], the radiation favours NW—NE. Of these two factors the wind is by far the most important one—53 % of the glacier-covered area is facing east as compared with only 4 % west, 20 % faces northeast and only 1 % northwest. A corresponding comparison for the radiation factor shows 10 % N versus 5 % S and 1 % NW versus 3 % SW[4].

However, where the two factors act together, i.e. in the sector N—NE—E, we find 80 per cent of the Swedish glaciers and 83 per cent of their area[5].

A SURVEY OF THE KEBNEKAJSE GLACIERS

Topographic description

As was stated above "Kebnekajse proper" has 17 glaciers (area 18.6 km²), 10 of which are larger than 0.5 km².

[1] Including dead ones, which were alive during the early part of the century.

[2] Only the 8 principal directions N, NE, E, SE etc. have been used—both for primary data and for the evaluation of the results.

[3] In some of the eastern parts of the Swedish high mountains a considerable amount of solid precipitation falls with easterly winds. This is, however, of little significance since the fallen snow is soon brought over on the eastern slopes by the predominant westerly winds of high velocity.

[4] In this study the influence of the bedrock structure has not been thoroughly investigated. However, it is worth mentioning that the dominant N-NE-E orientation is as well developed irrespective of the location of the glacier centres.

[5] It must be observed here that the *whole* area of every glacier listed as an E-glacier has been included in the figure under E in the table above even though considerable *parts* of the glacier may face other directions.

	N	NE	E	SE	S	SW	W	NW
Number of glaciers	23	67	99	22	15	3	4	4
Percentage of total	10	28	42	9	6	1	2	2
Area in km²	30	63	163	13	16	10	11	4
Percentage of total	10	20	53	4	5	3	4	1

Fig. 2. Most of the detailed research in Kebnekajse has been carried out on Storglaciären. The distinct end moraine is believed to show the glacier extent at the middle of the 18th century. The white summit in the background is the South Top of Kebnekajse, the highest point in Sweden. 25 July 1955.

These are:

a "Björlings[1] glaciär" (approx. 1.9 km^2) flows southeast and finally south from a firn area east of the South Top of Kebnekajse. The larger part of the firn area is above 1,650 m above sea level and the snout does not reach below the 1,450 m-contour.

b "Rabots[2] glaciär" is the largest one in Kebnekajse—approx. 4.5 km^2. It is nourished by snow accumulation in three cirques in the NW and W slopes of the main Kebnekajse ridge. From the rather low firn area, most of it between 1,400 and 1,550 m, the glacier flows with a very gentle slope down to 1,050 m.

c "Storglaciären". From a firn basin east of the Kebnekajse ridge three glaciers flow down into the Tarfala valley. The southernmost of these three, Storglaciären, is the largest one (3.2 km^2) and more glaciological research has been done on this glacier than on all other glaciers in Kebnekajse put together. Its main firn area lies between 1,500 and 1,650 m, the terminus is now at 1,100 m.

d "Isfallsglaciären" (the Icefall glacier) covers an area of 1.5 km^2. It is separated from Storglaciären and from Kebnepakteglaciären (the next one to the north) by wide ice- and firn-covered passes. As the name indicates, its slope is not as regular and gentle as that of Storglaciären or Rabots glaciär but the ice forms a proper ice-fall between 1,400 and 1,500 m. Its snout is at 1,180 m.

e "Kebnepakteglaciären". This is the northernmost of the described glacier triplets and also the smallest one (0.9 km^2). It slopes very steeply from the more than 1,600 m high snow reaches below Kebnepakte to the shore of the Tarfala lake at 1,160 m. The glacier is at present changing its shape rapidly because of a rock ridge which is now melting out and which has already cut off half the width of the glacier.

f "Enquists glaciär". On the leeward side (E) of Kuopertjåkko (about 1,970 m) there is a deep cirque which constitutes the main accumulation area of Enquists glaciär. The glacier (0.7 km^2) is also fed from another cirque (facing SW) further east and the tongue flows out in a SW-ly direction. The snout reaches down to approx. 1,350 m but is now difficult to define because of an abundant cover of surface moraine.

g "Norra Kaskasapakteglaciären". The ice from one large cirque east of Kuopertjåkko, and from one as large north of Kaskasapakte, flows out into one long glacier tongue reaching almost to the bottom (about 1,050 m) of the Kaskasavagge (vagge = valley in Lappish). Judged from the air photos its area is 1.5 km^2.

h "Sydöstra Kaskasatjåkkoglaciären" (0.6 km^2).

[1] J. A. Björling was the first Swede to climb Kebnekajse.

[2] Named after the French geographer Charles Rabot who made the first ascent of Kebnekajse.

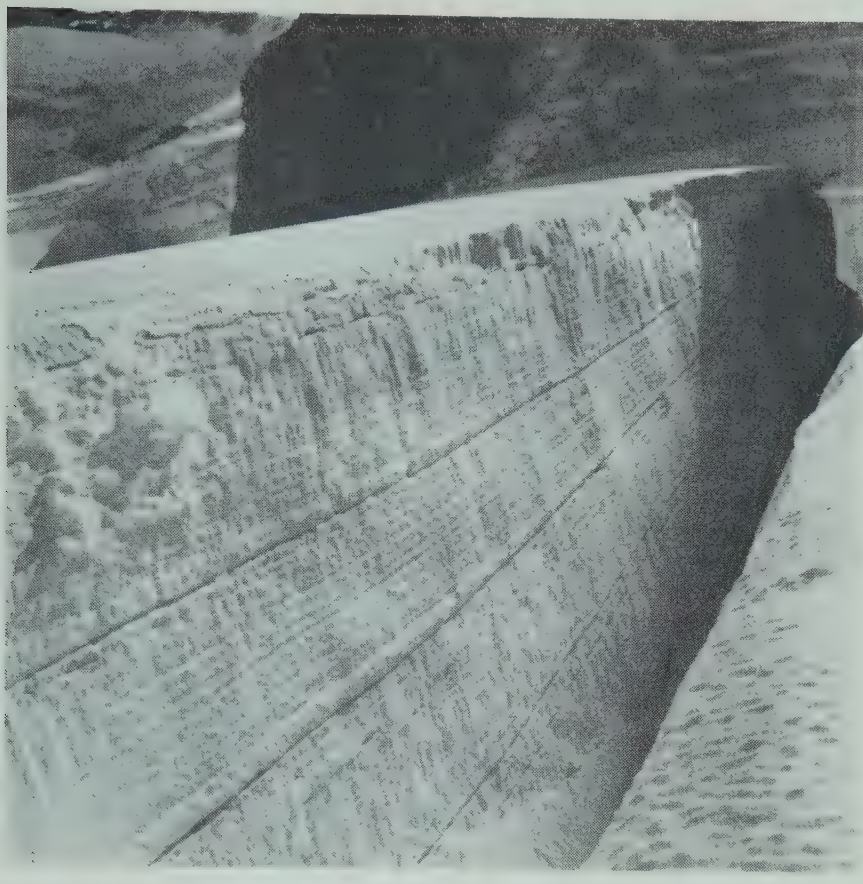


Fig. 3. A 3 m wide crevasse in the firn area of Storglaciären. The annual stratification is very distinct; the boundaries between separate snowfalls stand out as clearly.

The firn area (most of it above 1,550 m) is an accumulation leeward (E) of Kaskasatjåkko. From there the tongue flows south into the Tarfala valley reaching down to 1,320 m.

i "Nordöstra Kaskasatjåkkoglaciären" is of the same size as the previous one and separated from this by a narrow snow-free rock ridge. It flows north from an accumulation area east of Kaskasatjåkko.

j "Tarfalaglaciären" (1.1 km^2). This is the only large glacier in Kebnekajse which does not lie in a deep cirque. It occupies the smooth and gentle slope of Tarfalatjåkko just leeward of its highest point. From the main accumulation area at 1,600—1,700 m it reaches down to 1,440 m.

Regime investigations

The series of glaciological investigations at present going on in Kebnekajse was started in August, 1945, and since the spring of 1946 continuous observations of accumulation and ablation have been carried out on Storglaciären. Regime observations on other glaciers have been rather scarce but a few may be worth mentioning.

When the regime study was started in 1945 the research programme included work on both Storglaciären and Rabots glaciär. Arduous attempts were made to carry out such a comparative study but because of lack of means and man power a complete set of observations from Rabots glaciär could be obtained for only two years.

Before the ablation period started in 1946 (Schytt, 1947) the total winter accumulation (1945—46) on Rabots glaciär amounted to $3.5 \cdot 10^6$ m^3 of water ($0.8 \cdot 10^6$ m^3 per km^2). The snow depth increased regularly from 0.5 m close to the front to 2.5 m in the 1,550 m-region. During the summer the ablation removed $6.8 \cdot 10^6$ m^3 of water from the glacier ($1.6 \cdot 10^6$ m^3 per km^2) resulting in a net deficit of $3.3 \cdot 10^6$ m^3, almost as much as the total accumulation.

In an unpublished paper G. Östrem (1954) has reported his ablation measurements on Rabots glaciär in 1950. The observations covered only the period 12 july—9 Aug., and the ablation before and after these dates had to be computed from known ablation values at Storglaciären. Similarly the total accumulaton was mainly obtained through an indirect method. Östrem

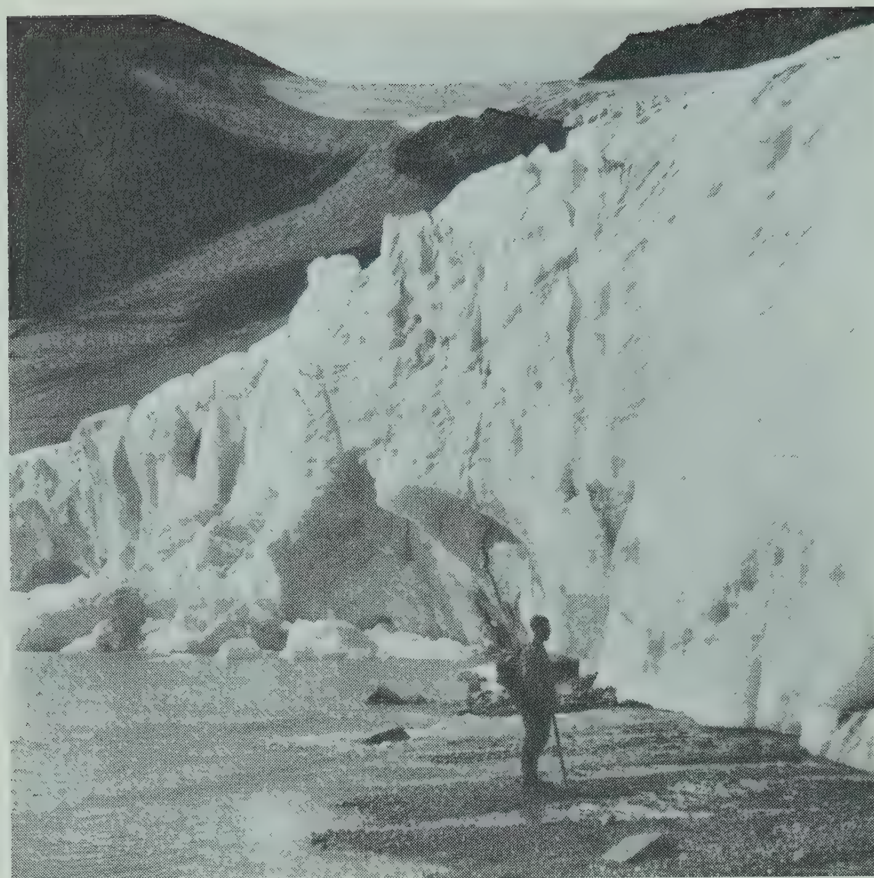


Fig. 4. The front of Kebnepakteglaciären in Aug. 1947. In 1945 the cliff was 25 m high—in 1959 there was hardly any cliff left.

states that the winter accumulation of 1949—1950 was $4.2 \cdot 10^6$ m^3 of water and the ablation during the summer of 1950 was $10.1 \cdot 10^6$ m^3. The deficit should thus have reached the high value of $5.9 \cdot 10^6$ m^3 (about $1.3 \cdot 10^6$ m^3 km^{-2}).

In accordance with these great deficits the net accumulation limit[1] (i.e. the lower limit of the accumulation area) was at 1,425 m in 1946 and and at 1,600 m in 1950. This means that in 1946 the accumulation area amounted to as little as 30 per cent of the area of the total glacier and that in 1950 practically the whole glacier (90 per cent of the area) showed a deficit. A balanced regime on Rabots glaciär would require an accumulation limit not higher than 1,350 m.

The ordinary tourist route to the top of Kebnekajse goes over Björlings glaciär. No systematic observations have been made here but it can safely be said that during a normal summer 80—90 per cent of the area below 1,625 m is deprived of its winter accumulation.

[1] We prefer to use "net accumulation limit" (normally abbreviated to "accumulation limit") instead of "firn limit" since the superimposed ice zone belongs to the accumulation area in spite of its position below the lower firn patches.

The few observations available from Tarfalaglaciären indicate that the accumulation area does not reach below 1,600 m.

On Storglaciären the regime observations were started in May 1946 (with a snow inventory for 1945—46) and have continued ever since. A complete snow inventory is carried out at the end of each accumulation season, i.e. late May or early June. All over the glacier a number of pits have to be dug through the winter snow in order to obtain representative values of snow density and snow depth. In the ablation area the winter accumulation can be determined very accurately by probing the snow depth with a steel rod and then converting the snow depth figures into water equivalents by using the density values obtained from the pits. The snow density at this time of the year usually averages 0.45 or slightly more. The probing is made along cross profiles with 50 m between each probing site and about 100 m between each line. Occasionally[2]

[2] Early in the season, before too many ice layers have formed in the winter snow, probing can be quite useful—after they have formed, probing should hardly be tried.

probing can be used even in the accumulation area but then only with several pits as check points. On the whole the accumulation figures from the higher parts of Storglaciären are based on pit observation, which means that they cannot be of the same accuracy as those from the ablation area—several profiles can be probed faster than one 4—5 m deep pit can be dug and evaluated. At the end of April 1960 a complete accumulation inventory by core drilling will be made in the higher reaches of Storglaciären. Then the last winter's snow will still be almost free of ice layers and it will be easy to establish the position of the previous summer surface.[1] In this way it should be possible to obtain such a great number of observations even in the accumulation area that the amount of snow here should be known as accurately as that in the ablation area.

The ablation is normally measured with the ordinary stake method. This works very well when the glacier ice is exposed, provided that the holes into which the stakes (1 inch Al-tubes) are planted are deep enough[2]. When snow ablation is measured the difficulties become greater. The settling of the snow and the refreezing of the melt water make observations complicated. The ablation over a certain period at a certain place in the accumulation area can best be obtained from two separate "accumulation measurements". This means, that the total water equivalent of, for example, the last winter's snow has to be measured in pits at the beginning and the end of the period. This cannot be done at short intervals at several places; therefore the stake method has to be used. However, the results must be corrected for such density changes as are observed in a limited number of pits dug throughout the summer[3].

Quantitative data:

The total accumulation varies appreciably and rather irregularly over the glacier. 1.5—2 m of water can be given as a normal value for the higher parts (above 1,450 m) and 0.5—1.0 m for the lower parts (below 1,400 m). However, because of the local topography, variations are considerable even in one single cross profile. On 14 May 1959, for example, one large patch of ice was exposed on the south side of the glacier at 1,275 m elevation, while on the north side and at the same altitude the snow depth was 2 m. Even though the total accumulation varies greatly from one year to another (see table 1) the snow distribution shows approximately the same picture.

The ablation is more dependent upon height above sea level than upon local topography. It is true that the slope of the surface, the closeness

[1] The accumulation after the day of core drilling can be measured by the stake method or even better by staining the surface and then by digging shallow pits later in the summer.

[2] In 1946 it was observed that stakes planted in shallow holes (<100 cm) always showed too small ablation. Three series of observations gave the following errors: —40 %, —28 % and —25 %. Planted in deep holes (normally about 300 cm) they register the true ablation.

[3] In 1946 comparative studies between the stake method and the pit method showed that the stake method could give up to 44 % too small ablation figures.

Table 1. Regione of Storglaciären 1945—1958.

Regime year	Accumulation $10^6 m^3$	Ablation $10^6 m^3$	Deficit — surplus + $10^6 m^3$	Accum. limit m above sea level
1945—1946	3.5	7.0	— 3.5	1,480
1946—1947	3.2	9.6	— 6.4	1,600
1947—1948	4.5	4.5	+ 0.0	1,400
1948—1949	6.9	4.1	+ 2.8	1,410
1949—1950	4.4	8.4	— 4.0	1,550
1950—1951	2.5	4.5	— 2.0	1,500
1951—1952	2.7	3.2	— 0.5	1,450
1952—1953	6.0	8.5	— 2.5	
1953—1954	3.5	6.5	— 3.0	
1954—1955	5.0	5.5	— 0.5	
1955—1956	4.0	5.5	— 1.5	
1956—1957	5.0	6.0	— 1.0	
1957—1958	4.5	6.5	— 2.0	

to exposed rock walls, etc. exert an influence upon the rate of melting, but still, if the ablation is plotted as a function of altitude, a smooth curve is always obtained. Close to the front an ablation of 4—6 m of water can be expected during a normal summer, at 1,300 m it has dropped to 2—4 m, at 1,400 m to 1.5—3.5 m and at 1,600 m to 1—2 m.

Table 1 shows the total accumulation, total ablation and the regime balance of Storglaciären during the years 1945—1958, i.e. 13 regime years. The mean accumulation over this period was $4.3 \cdot 10^6$ m^3 year^{-1} ($1.3 \cdot 10^6$ m^3 km^{-2} year^{-1}). The deviations from the mean were considerable —the minimum accumulation value (1950—51) was $2.5 \cdot 10^6$ m^3 and the maximum was $6.9 \cdot 10^6$ m^3. The high value of $6.0 \cdot 10^6$ m^3 followed immediately after the low of $2.7 \cdot 10^6$. The standard deviation amounts to $\pm$ $1.2 \cdot 10^6$ m^3 (27 % of mean value).

The ablation varies even more. The average total ablation for the 13 year-period was $6.1 \cdot 10^6$ m^3 year^{-1} ($1.8 \cdot 10^6$ m^3 km^{-2} year^{-1}); the maximum was $9.6 \cdot 10^6$ m^3 in 1947 and the minimum was $3.2 \cdot 10^6$ m^3 in 1952. The standard deviation was $\pm$ $1.8 \cdot 10^6$ m^3 (29 % of mean value).

Finally, the net regime was balanced in 1947—1948, positive in 1948—1949 and negative all the other years. In 1947 the deficit was twice as large as the total accumulation. Even at that time it was evident that the summer of 1947 was very unusual. Just a few small patches of last winter's snow remained on the very highest parts of all the Kebnekajse glaciers at the beginning of September, practically all "perennial" snowfields melted off completely, and our figures show that the average thinning of the whole glacier, the accumulation area included, was more than 2 m (1.9 m of water). During a normal year the average thinning amounts to about two thirds of a metre ($1.9 \cdot 10^6$ m^3 or $0.6 \cdot 10^6$ m^3 km^{-2})[1].

It may be of interest to observe, that during an average summer this fairly small glacier contributes $6.1 \cdot 10^6$ m^3 of water to the water resources of the river Kalix älv; if this rate of ablation is accepted as a rough mean for the whole glacier covered area of Sweden, our rivers would receive approximately $600 \cdot 10^6$ m^3 of water per year from glaciers. This has naturally some favourable effect upon the water storage for hydroelectric power since the glaciers add most of their water to the rivers after the main snow melt. This great run-off has, however, used up some of the capital. During 1945—1958 Storglaciären had a net loss of volume corresponding to $24 \cdot 10^6$ m^3 of water ($7.3 \cdot 10^6$ m^3 km^{-2}). Each year the Swedish glaciers[2] would, according to this, lose about $175 \cdot 10^6$ m^3 of water because of the present climatic fluctuation. Should the present trend change and the annual deficit turn into as large an annual surplus for the glaciers, the supply of glacier melt water to our rivers would drop to less than half of its present value.

The dividing line between the accumulation (surplus) area and the ablation (deficit) area has, in our terminology, been called (net) accumulation limit. It corresponds strictly to the "old" concept "firn limit" if this is defined as "the line at which the year's total accumulation equals its total ablation" (Ahlmann, 1948, p.41, line 5), but not if defined as "the highest level on a glacier to which a winter snow cover recedes during the following season" (Ahlmann, 1948, p.41, lines 3—4). A special study on Storglaciären (Schytt, 1949, p.227) showed that the winter snow receded a distance of from 10 to 200 m beyond the line where ablation equaled accumulation; the intermediate area was covered with a sheet of superimposed ice.

Since the altitude of the accumulation limit is dependent on both accumulation and ablation, it varies considerably from one year to another. The data available today (a detailed analysis will be published by E. Woxnerud) indicate a linear relationship between the altitude of the accumulation limit and the regime balance. During a balanced year (e.g. 1947—1948) the accumulation limit is found at about 1,425 m and for every $1.0 \cdot 10^6$ m^3 of deficit it climbs about 25 m higher up glacier.

Movement studies on Storglaciären

Woxnerud has carried out a long term study of the ice movement of Storglaciären. As a general rule it can be stated that the maximum movement (about 13—15 m year^{-1}) is found along the middle line of the glacier. The lines connecting

[1] A complete account of the regime investigations is being worked out by Dr E. Woxnerud, who has kindly put the data in Table 1 at the present author's disposal for this short survey.

[2] We here assume that the conditions of Storglaciären can approximately represent all Swedish glaciers.

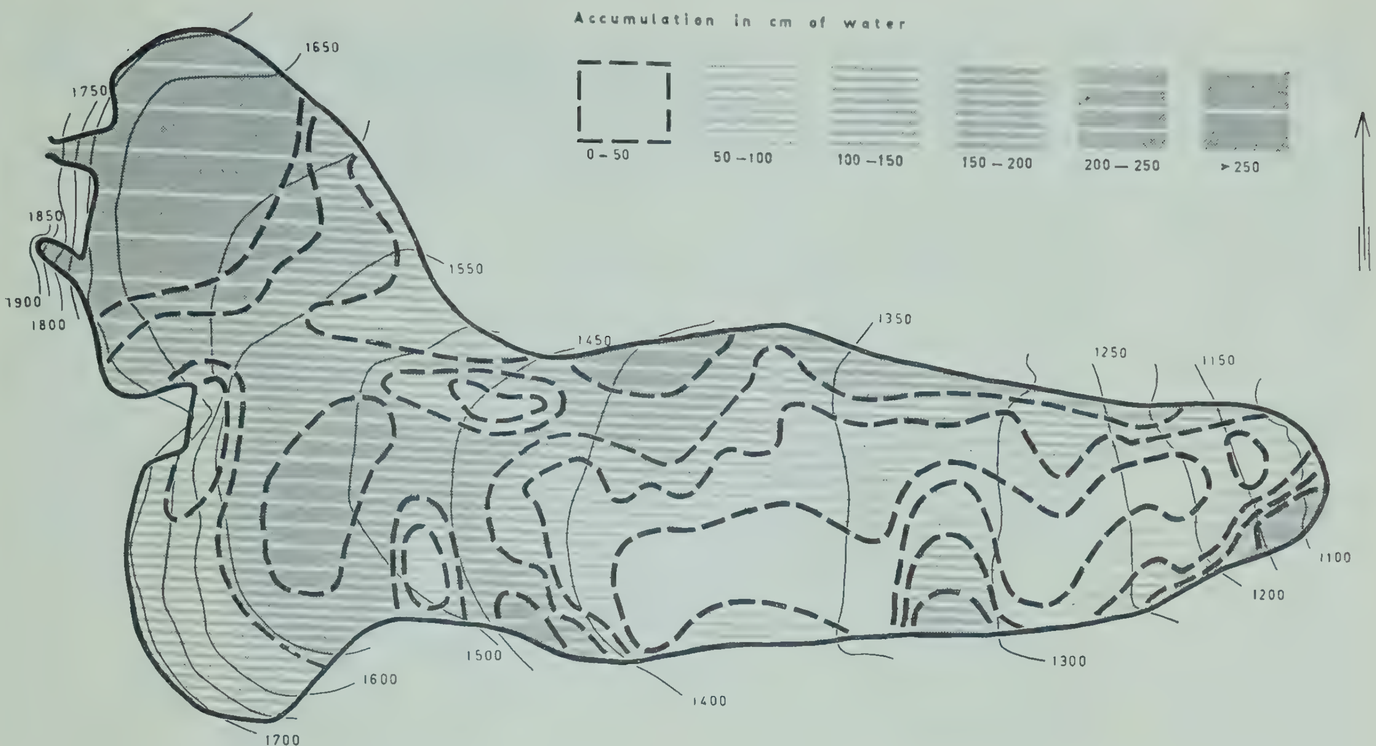


Fig. 5. Map showing the distribution of accumulation on Storglaciären during a "normal" year. Map prepared by E. Woxnerud. Scale approx. 1 : 27 000.

places with the same rate of movement run parallel to the "shores"; the 1 cm day^{-1}-line runs 50—150 m from the edges, the 2 cm day^{-1}-line 150—250 m from the edges and the 3 cm day^{-1}-line another 100 m further in. The movement pattern is fairly symmetrical except close to the front where the maximum movement is displaced towards the north (left). This is due to an ice-covered rock ridge under the southern part of the glacier at an altitude of 1,200 m. The ridge could hardly be detected today, was it not for the movement pattern. However, 50 years ago, or even less, when the glacier reached close to its outmost terminal moraine, there was an ice-fall at the 1,200 m level. Today, ice-thickness is too small and movement too slow to cause an ice-fall.

Woxnerud has found the summer (June—Aug.) movement on the glacier tongue to be about 40 % higher than the winter (Sept.—May) movement; in the upper reaches of the glacier summer and winter movement are nearly the same, and in the very highest parts of the accumulation area winter movement even seems to predominate.

Moraine features

Erik Bergström has made a close study of the moraines at 21 glaciers in the Kebnekajse area (Bergström, 1954 and unpublished reports). Some of the glaciers have only one terminal moraine, most of them have three or four and some have up to twelve each. He states that the terminal moraines have been formed in three different ways: (1) an advancing glacier may act as a bull-dozer and push forward surface material from a previously deposited moraine cover, (2) during a stagnation period surface moraine will fall down and collect along the front, and (3) debris will be carried out to the front as ground moraine or along shear planes. The larger terminal moraines in Kebnekajse have been built up during stagnations and by both surface material and debris from the inner parts of the glaciers.

The lateral moraines studied do not occur in such great numbers as the terminal moraines and are not so well separated from each other. An inner lateral moraine is normally deposited on top of its outer neighbour and they can be identified as two units only by the two narrow ridges along the crest.

Storglaciären

Three terminal moraines can be seen. The outer one consists of huge blocks (5—10 m diam. is common) and has very little fine material. The fines have been flushed away by marginal drainage

at a time when the glacier dammed up the whole valley and all the water from the other glaciers had to pass along the front of Storglaciären. Two deep (up to 8 m) drainage channels just outside the southeastern corner of this terminal moraine also bear witness of a considerable water erosion at the time of this maximum extension. The two other terminal moraines are not so well developed, they consist of smaller blocks and they run about 20 m and 40 m, respectively, inside the outer one.

The lateral moraines are much more impressive than the marginal ones. On the south side the moraine ridge comes out of the ice at the 1,375 m level and can then be followed for over 2 km down to the bottom of the valley. The ridge, consisting of several more or less well separated units, rises 20—30 m above the ice. The block content is very high, the blocks are angular with sharp edges and it is obvious that the bulk of the material has fallen down from the steep walls of Kebnetjåkko and been very little affected during the transportation within the ice or on the top of it. The further down slope the material is observed, the higher is the content of fines.

On the north side there is abundant accumulation along the foot of Södra Klippberget (Southern Rock) and the huge lateral moraine disappears underneath the snow at 1,350 m. As in the case of the southern moraine, the block content is high in the upper parts and diminishes downwards. Several stagnations have built up a rather complicated ridge pattern.

Just outside the present front of the glacier there is a large area which has been interpreted as moraine cover on top of dead-ice. The debris started to melt out as a shear plane moraine in 1946 and the formation has continued ever since.

The deposit consists mainly of fine material, saturated with water and quite unconsolidated. It is easily eroded by water action, and though enclosed ice has been looked for in several eroded sections none has ever been found. In August 1959 geo-electric measurements indicated that all previous dead-ice (observed in 1946 and the years after) had apparently melted away. An investigation of this large accumulation of fine, previously frozen but now water-saturated, debris will soon be initiated. The main investigation will be devoted to particle orientation, structure and stratification.

Dead-ice certainly exists outside the present northeast corner of the glacier and as a core in the lateral moraines. However, to what extent these moraines are ice-cored is at the moment quite unknown. Research along these lines was not started until 1959.

Isfallsglaciären

This glacier has more interesting moraine features than any other glacier in the Tarfala valley. Three terminal moraines are very well developed, but in contrast to the conditions at Storglaciären, the outer one is the lowest and the inner one the highest. Sand and gravel dominate on the crests and on the inner sides of the moraine ridges while the block content is particularly high on the outside slopes.

The lateral moraines are short, about 500 m, but higher than those of Storglaciären (up to 50 m). On the north side several well-separated ridges show that a great number of stagnations have occurred during the general retreat since the post-glacial maximum. Frequent slumping takes place in the southern slope of the northern moraine and clear, "black" dead-ice is often exposed. Much is not known about the amount of dead-ice in these moraines, but rather likely the debris-covered ice rests like a fairly thin sheet on the inner side of the previous ridge. This, in its turn, may hide another slab of ice and the whole series of lateral ridges may contain several thin bodies of dead-ice, all dipping steeply towards the glacier. The only seismic profile hitherto available showed the existence of an 8 to 10 m thick ice core hidden under 1.5—2 m of moraine[1].

Most of the area between the ice front and the innermost lateral moraine looks very much like a newly-ploughed field. High ridges and deep furrows run straight out from the ice edge and several of them can be traced all the way up to the crest of the 15 m high terminal moraine—250 m from the present ice front. The ridged pattern has been called "fluted moraine" and this particular occurrence has been described in Hoppe and Schytt, 1953.

All ridges run parallel to the ice movement; the larger ones are 30—40 cm high and they are

[1] Evidently dead-ice can be found in very old moraine ridges. E. Bergström and A. Olsen (unpublished reports) found exposed glacier ice in a 100 years old terminal moraine in front of Norra Kaskasapakteglaciären 4 km further north.

6. Fluted moraine in front of Isfallsglaciären. July 1955.

1—1.5 m apart. There are, however, a great number which are only a few cm high and spaced less than 25 cm apart. Close to the ice front the ridges, except for the very surface layer, stay frozen but further out they are loose and water-saturated and one can easily sink a foot or so in the mortar-like mass. Still further away from the ice, especially on the slopes of the terminal moraine, the ridges are dry and "solid".

In 1949 and 1952 trenches were dug in the ice 5 m above where some distinct ridges disappeared under the front. All ridges were found to continue inwards. They were all twice as high under the ice as they were outside—a result of the high ice content in the submerged ridges. This ice, which was observed as lenses and thin laminae often parallel to the sides of the ridge, was not glacier ice but frozen melt-water with long columnar crystals. The material was quite unsorted and contained all grades from clay to boulders; all stones were well polished. No melt-water was found in or between the ridges, everything was frozen solid. Three of the examined ridges showed a symmetrical cross-section, the fourth was tilting over towards the north.

At all places, where fluted moraine has been observed by the author so far, there has been an abundance of fine material. Several ridges start from a large boulder, which has apparently been lying in the same position for a long time, as judged from the consistent striation. It should also be observed that the glaciers in Kebnekajse have a temperature slightly below freezing at the end of the summer season. This is true at least down to 10 m. Consequently the ground moraine under the ice front stays frozen the year round; how far in is not yet known.

All observations indicate that fluted moraine is an *accumulation* phenomenon but not related to running water. The ridges are more probably caused by a continuous supply of more or less fluid ground moraine which is pressed up in cavities formed behind (on leeward side of) boulders fixed solidly in the substratum. It has been observed that the ridges keep an astonishingly constant height even far out from the front (i.e. as soon as all the ice has thawed out), and this indicates that the cavity filling can hardly stay in fluid state after it has been pressed up.

Fig. 7. The snout of Storglaciären and its terminal moraines seen from the west. 25 July 1955.

If it did, the ice pressure would soon lower the roof over the ridge, and we would get a moraine-tail behind the boulder—not a several hundred metres long ridge, which can even go up-hill. It is thus believed that the fluid moraine, which is originally at pressure melting point, is frozen to the basal ice because of the release of pressure when it is pressed up in the cavity. It is then carried along with the ice, and fresh debris, from the sides and coming down with the moving ice, is continuously pressed up in the lee of the boulder and added to the previously frozen cavity filling. In this way the ridge, being built up from behind, is carried forward as a part of the moving glacier until it reaches the zone where the low winter temperatures extend through the ice into the ground moraine. There, the ridge freezes to the substratum and can no longer move with the ice, which from now on must slide over the ridge. Since there is now a solidly frozen moraine ridge between the initiating boulder and the outer "stranded" end, no more cavity can form behind the boulder, and the formation of the ridge has come to an end.

Such an explanation accounts for the parallelism to the direction of movement as well as for the constant height and considerable length of individual ridges.

If it is true that fluted moraine is caused by water-saturated debris being pressed up in cavities behind boulders, this morphological feature can be used as a kind of "bottom thermometer". When fluted moraine exists outside a glacier the temperature at the glacier bed must be at pressure melting point. This would be a useful instrument for glaciological research since so very little is known about bottom conditions. It may be added, that during a flight along the west coast of West-Spitsbergen Hoppe and Schytt observed fluted moraine in front of at least one large glacier reaching almost down to the sea level.

Glacier variations

After the retreat of the last inland ice sheet the climate of the Swedish mountains was considerably warmer than it is today. If glaciers at all existed they must have been confined to the very highest mountain massifs, viz. Sarek and Kebnekajse. Because of the climatic change which occurred at about 500—600 B.C. many new glaciers formed and already existing glaciers advanced. It is believed that there was a glacier maximum during the last centuries B.C. and a minimum during the first centuries A.D. (Bergström, 1954). Our recent glaciers apparently reached their maximum extent during the middle —or maybe early part—of the 18th century when

most of the glaciers built up their outermost terminal moraines. During the last 200 years there has been a general glacier retreat with secondary maxima at about 1810 and 1910 and a very rapid retreat which started about 1920.

Storglaciären

The outermost terminal moraine is according to E. Bergström from about 1750[1]. In 1897 the glacier was visited by Nordgren and Rönnholm (Svenonius, 1910) who found the ice front only 31.5 m inside one of the huge blocks (R_2) of the moraine ridge. Between 1897 and 1908 there had been a slight advance (distance to R_2 only 26 m in 1908) and photographs from 1922 show that very small changes had taken place until then. During the following years a very rapid retreat set in as the following table shows:

Year	Distance to R_2
1897	31.5 m
1908	26 m
1922	30—40 m?
1929	95 m
1945	230 m
1950	320 m
1955	385 m
1959	405 m

The maximum rate of retreat seems to have been reached at the middle of the 40's with averages around 20 m/year. This retreat is of course a result only of the thinning-out of the glacier, a factor which can be computed from four different glacier maps made in 1922, 1946, 1949 and 1959, respectively.

The moraine just outside the 1946 ice front was in 1922 covered with a 75 m thick ice sheet. At 1,200 m (above sea level) the glacier had become about 40 m and at 1,350 m about 30 m thinner in the period 1922—1946. At the 1959 ice front the glacier of 1949 was 20 m thick.

Between 1922 and 1946 Storglaciären had, according to the maps, a net loss of about 70 million m^3 or $2.6 \cdot 10^6$ m^3 of water per year. Our direct observations during the years 1945 to 1958 gave an annual net loss of $1.9 \cdot 10^6$ m^3.

Isfallsglaciären

Old photographs (1903, 1908, 1910 and 1922) and reported distance between the ice front and a conspicuous block, R_4, in 1897 and 1910 show that the ice advanced 40—50 m just before the turn of the century and that no major changes took place between 1903 and the early 20's. In the 18th century, however, the glacier was much more advanced, which is shown by a terminal moraine 100—150 m outside the ice front of 1910 and 1920.

Old photographs show that the ice before the last major retreat (i.e. until about 1925) was dammed up by the high terminal moraine. The actual front was situated well on the outside of the crest, and it is interesting to see how a glacier for many years can over-ride a terminal moraine without destroying it. It is still an open question when this high (15—20 m) moraine ridge was built up.

Bergström's lichenological studies indicate that during a stagnation around 1850 the ice front stood only 35 m inside the "1750-moraine". It therefore seems most probable that the high terminal moraine (100—150 m inside the "1750-moraine") is at least older than the 1850-stagnation, perhaps even older than the outermost terminal moraine.

This somewhat peculiar chronology throws some doubt upon other datings of terminal moraines, especially since another terminal moraine has been observed to be melting out of the snout of Rabots glaciär. When the present investigations started in 1945 only one large terminal moraine could be seen, a few years later a transverse ridge started to melt out at the ice edge.

During recent years Isfallsglaciären has retreated considerably. The front is now standing about 400 m inside the "1750-moraine" and about 250 m from the front position of the early part of this century.

Kebnepakteglaciären

The retreat of this glacier has had to be studied from photographs only. It amounts to between 200 and 300 m since about 1925. 30 years ago calving took place along the whole ice front, which in 1945 was 25 m high. The ice edge has now climbed up on land in almost its full length, and it will soon be possible to walk around the lake shore without setting foot on ice.

Toppglaciären

The highest summit in Sweden is the South Top of Kebnekajse—capped by a small glacier. On

[1] Recent lichen studies by LARSSON and LOGEVALL indicate a greater age than 200 years (private communication).

3—*910100 Sep.*

the official topographic maps (1:200,000) the summit height is given as 2,123 m, a value obtained through measurements on 14 July 1902. These observations were repeated in Aug., 1947 and Aug., 1948 by V. Schytt and in July, 1950 by E. Woxnerud (Woxnerud, 1951). They found that because of the general glacier retreat even the top glacier had decreased in thickness. The differences as compared with the 1902 value were: —9 m, —8 m and —4 m for the three sets of observations. It should be added that the ablation during the summer of 1947 was the highest ever measured, and that the 1950-observations were carried out 1 ½ months earlier than those in 1947 and 1948—even the exceptionally great accumulation of the winter 1948—1949 may still have exerted its influence.

It is obviously impossible to give an accurate value of the height of Kebnekajse—since it varies both annually and seasonally—but 2,115 m can be used as a good mean value.

All the other glaciers in "Kebnekajse proper" have been studied at various occasions, especially by E. Bergström. They have responded to the present climatic fluctuation in very much the same way as Storglaciären and Isfallsglaciären. It is interesting to observe that one group of glaciers either reached their maximum extent (such as Rabots glaciär) or at least very close to the 1750-moraine (e.g. Storglaciären) in the early part of our century, while another group (e.g. Isfallsglaciären) stood at the inner one of three (occasionally four) terminal moraines spread out over about 100—150 m.

The migration of the vegetation

It has been observed that the tree-line during recent years has climbed higher up on the mountain slopes and that several new species have been found at high altitude. It is of greatest glaciological interest to observe how the vegetation is slowly invading the large moraine areas which 40 years ago were covered with ice.

Erik Bergström carried out extensive investigations at 21 Kebnekajse glaciers in 1948. He was able to distinguish between four different vegetation zones mainly using lichens as characteristic plants: *a.* an outer zone which is "completely" covered with a great variety of plants and where the green Rhizocarpon (e.g. geographicum and oreites) is the dominant lichen and occurs in large sizes, *b.* one zone where numerous specimens of Rhizocarpon are found in a dense carpet of Umbilicaria (proposhidia and cylindrica), *c.* a "black zone" where Umbilicaria is entirely dominant and Rhizocarpon rare, *d.* an inner "light zone" where lichens cover only an insignificant part of the moraine surface. Bergström could show that the outer zone (a) had not been covered with ice for a very long time and was separated from the next zone (b) by the "1750-moraine". Between this moraine and the middle one of the usually existing three main terminal moraines comes zone b, which should represent the area uncovered by ice between the middle of the 18th and the beginning of the 19th century (given tentatively by Bergström as 1735 $\pm$ 50 years and 1807, respectively). The "black zone" of only Umbilicaria was laid bare of ice between the beginning of the 19th and the early part of the 20th century, while the light zone shows the area from where the glaciers have retreated since about 1920.

More lichen studies were carried out in Kebnekajse in 1958 and 1959 by B. Larsson and B. Logewall (unpublished report). They used a quantitative method partly adopted from Beschel (1957), and they found great differences in lichen diameter between the inner terminal moraine (reached or even covered in 1910—1920) and the old ridges from the 18th century[1]. Since the recent retreat started to lay the inner moraine bare (1910—1920) the largest Rhizocarpons have grown to an average diameter of 4 mm, Umbilicaria proposhidia 24 mm and Umbilicaria cylindrica 22 mm[2]. On the outermost, about 175 years older, moraine the average size of large Rhizocarpons is 50 mm, U. proposhidia 37 mm and U. cylindrica 30 mm[3]. Bergström's observations have thus received a convincing quantitative support. More work is going on along these lines in the search for even older terminal moraines, e.g. from late-glacial times, and a small number of such ones have already been found.

A few remarks about the deglaciation of the area

The old concept of an ice-free mountain range and a great mass of ice melting down over the,

[1] For each sampling area (25 m^2) the diameter of the five largest species of different lichens were measured and the average was computed.

[2] These are averages for Isfallsglaciären, Storglaciären, S. E. Kaskasatjåkkoglaciären and Björlings glaciär.

[3] Averages from Isfallsgl. and Storglaciären.

at present, forested country to the east has now been definitely abandoned. There is ample evidence that the ice retreated towards some of the high mountain massifs that acted as local culmination centres during the last stages of the deglaciation.

Kebnekajse was one of these centres, and the very conspicuous lateral drainage channels on several valley slopes prove this without any doubt. When approaching Kebnekajse from the east one can notice very well developed systems of lateral channels along the slopes south of the lake Paittasjärvi, on Larkimvare south of the Lappish village Nikkaluokta and along both sides of the main west-east valley, Ladtjovagge. Several of the channels are very long and in their easternmost ends some of them turn sharply down towards the bottom of the valley thus showing not only the side but also part of the front of the retreating glacier.

Detailed investigations of these drainage channels are going on and in 1959 one part of the mountain Larkimvare was surveyed by L. Lundin and L. Skålander. They found (unpublished report) that the average gradient in the upper reaches of the channels was about 1.6 m in 100 m and that as much as 2.3 m in 100 m could be observed farther down—but still high above the valley floor. The vertical distance between the channels varied between 5 and 14 m and it is not believed that this can be taken as a value of the annual net melting of the ice (compare Schytt, 1956).

Farther west, at the mouth of the Tarfalavagge, G. Adolfson and P. Palm leveled a number of lateral drainage channels in 1959. At an altitude of 900 m above sea level the gradients were around 2.3 m per 100 m.

Through an intensive study of such channels as well as subglacial eskers and chutes, various lateral accumulations, striae, drumlins etc. it will be possible to obtain a very detailed picture of the deglaciation of a high mountain massif, i.e. of the disappearance of one of the very last remnants of the last Scandinavian inland ice sheet.

References

AHLMANN, HANS W:SON. 1948. Glaciological research on the north Atlantic coasts. R. G. S. Research Series, No 1. London. Roy. Geogr. Soc.

BERGSTRÖM, ERIK. 1949. Undersökningar av några Kebnekajse-glaciärer speciellt med hänsyn till deras horisontella och vertikala tillbakagång (unpubl. report).

BERGSTRÖM, ERIK. 1954. Studies of the variations in size of Swedish glaciers in recent centuries. U. G. G. I. Ass. Int. d'Hydrologie. Publ. no 39 (pp. 356—366).

BESCHEL, R. 1957. Lichenometrie im Gletscher-Vorfeld. Jahrbuch 1957 d. Vereins zum Schutze d. Alpenpflanzen und — Tiere.

ENQUIST, FREDRIK. 1916. Der Einfluss des Windes auf die Verteilung der Gletscher. Bull. Geol. Inst. Upsala. Vol. XIV.

HOPPE, GUNNAR and SCHYTT, VALTER. 1953. Some observations on fluted moraine surfaces. Geogr. Annaler 1953 (pp. 105—115).

LUNDQVIST, G. 1948. De svenska fjällens natur. Stockholm.

OLSEN, ASBJÖRN. 1949. Några iakttagelser angående dödisar och ackumulationsformer gjorda i samband med moränundersökningarna i Kebnekajsemassivet sommaren 1948 (unpubl. report).

SCHYTT, VALTER. 1947. Glaciologiska arbeten i Kebnekajse. Ymer 1947 (pp. 18—42).

—1949. Re-freezing of the melt-water on the surface of glacier ice. Geogr. Annaler 1949 (pp. 222—227).

—1956. Lateral drainage channels along the northern side of the Moltke Glacier, northwest Greenland. Geogr. Annaler 1956 (pp. 64—77).

SVENONIUS, F. 1910. see Die Gletscher Schwedens . . .

WALLÉN, C. C. 1949. Glacial-meteorological investigations on the Kårsa Glacier in Swedish Lappland 1942—1948. Geogr. Annaler 1948 (pp 451—672).

WOXNERUD, E. 1951. Det lokala triangelnätets i Kebnekajse anslutning till riksnätet. Syd- och Nord-topparnas höjd över havet. Geogr. Annaler 1951 (pp. 131—143).

ÖSTREM, GUNNAR. 1954. Materialhusholdningen på Rabots bre 1945—46 og 1949—50 (unpubl. report).

"Die Gletscher Schwedens im Jahre 1908". Sveriges Geol. Undersökning. Ser. Ca. No 5. 1910.

APPENDIX

ICE MELTING UNDER A THIN LAYER OF MORAINE, AND THE EXISTENCE OF ICE CORES IN MORAINE RIDGES

BY GUNNAR ÖSTREM

Department of Geography, University of Stockholm

Introduction

It has been known for a long time that some of the moraine ridges in the Tarfala valley are ice-cored. As this ice melts away, the moraine material which lies on the ice slides down and protects the lower part of the ice core from further melting. Meanwhile in the upper part, the process will continue, especially during warm, sunny days.

In order to get information about a moraine layer's influence on ice melting, some ice melting experiments under a thin layer of moraine cover were carried out during the summer of 1956.

Experiments with ice melting

On Isfallsglaciären, sand and gravel was placed on the clear glacier ice in test fields about 2 m^2 in area. The melting of the ice under the sand and gravel cover was measured by means of bamboo stakes. Control measurements on the uncovered glacier ice were also carried out by means of bamboo stakes.

By using moraine material of different grain size, and by making the test covers of different thicknesses, the experimental conditions could be varied.

Meteorological factors were observed at the nearby research station in Tarfala.

The temperature distribution between the upper and the lower level of the sand layers could be observed by means of thermistors. By extrapolation from the measured temperatures at different depths, the mean temperature on the upper sand surface during the measuring period (10/7—5/8 1956) could be computed (it was $+9°C$). As the mean air temperature during the same period was only $+5.4°$ C, the warming of the sand and gravel layers must come to a great extent from the insolation. The absorbed amount of energy will be distributed among the following:

1. Outgoing (longwave) radiation.
2. Energy loss to the air by convection and conduction.
3. Energy loss through evaporation of melt water which has risen in the sand and gravel by capillary action.
4. Melting of glacier ice.

It was evident that the distribution of energy among these four factors varied with grain size and thickness of dirt layer.

To get more precise data on ice cores in moraine ridges, special attention was given to the effects of grain sizes and layer thicknesses upon the rate of melting. It was evident from direct measurement, that the uncovered ice melted with a mean rate = 4.5 cm/day. This rate decreased when the cover was more than about ½ cm thick. For example, under a 6 cm thick cover the mean melting speed was 3 cm/day; under 20 cm cover it was less than 1 cm/day. (Fig. 1).

[Not only will the melting be slower under a moraine cover, but also the ablation period will be shorter for the covered ice.]

Under really thin layers, the melting speed will accelerate, but as exact measurements are difficult to obtain because of melt water erosion, this part of the graph has been dotted only. The

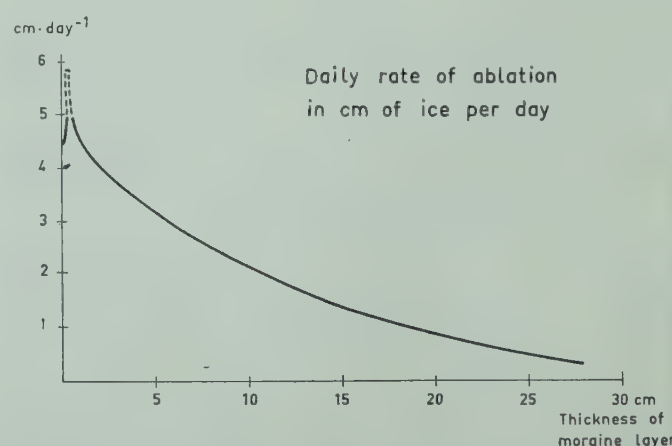


Fig. 1. The daily rate of ablation during the measuring period 10/7—5/8 1956 when the normal ablation of exposed glacier ice was 4.5 cm/day. Computed from measured results at the test sites on Isfallsglaciären.

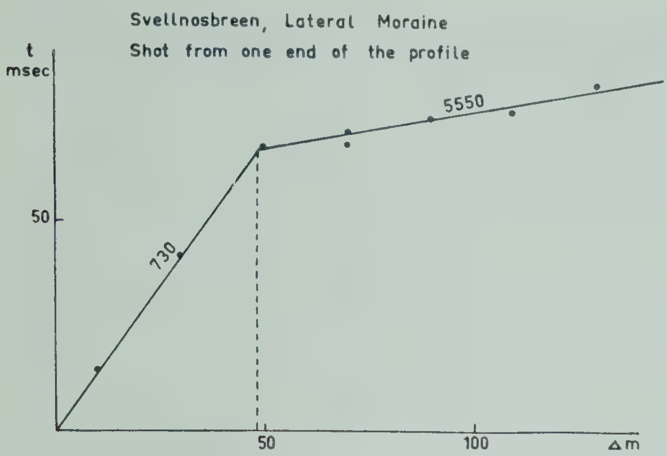


Fig. 2. Travel-time graph made from seismic shots on the lateral moraine ridge at Svellnosbreen. No ice core can be determined. The moraine material lies on bed rock.

maximum point of the curve has been computed from the known value of incoming radiation and by setting the different losses (to evaporation, etc.) as great as the losses would be from an uncovered ice surface.

Since, as seen from fig. 1, the curve appears to approach the X-axis as an asymptote, it is difficult to determine the thickness of moraine cover which should be great enough to permanently preserve ice from all melting. (It will be the point where the curve meets the X-axis). To get information about this limit value, it is necessary to measure the thickness of moraine cover on old ice deposits. This means principally on old moraine ridges.

Two methods have been used in these investigations: the seismic and the electric resistivity methods.

Localisation of ice cores in moraine ridges

In March, 1959, measurements were made at different places in the Tarfala valley, as well as on the lateral moraine at Isfallsglaciären. These measurements were made with a 12-channel seismic refraction instrument.

The thickness of the ice core was measured as 8—10 meters, but since the moraine cover was quite thin, the seismic method did not give information about this layer's thickness. Consequently, it was necessary to use other methods.

Efforts have been made to dig in the unconsolidated material on the moraine ridge, but without results.

Therefore, in August, 1959, several electrical earth resistivity measurements were made in the valley. Both the so-called Wenner (See Dobrin, 1952, p. 295) and the Schlumberger electrode configurations (See Lasfargues, 1957, p. 66) were used. At the same point on the lateral moraine of Isfallsglaciären where seismic measurements had been made, it could be shown that the moraine cover was only 1.8—2.6 metres. (The difference is the result of different computing methods.) Under this layer was an insulating body, the resistivity of which indicated that it must be ice. The thickness of this ice core could not be computed from the electric resistivity measurements.

Based upon these parallel experiments in the Tarfala valley, more investigations were made in the Jotunheimen area in Norway during October, 1959, and an additional survey in the Kebnekajse massif is planned for the summer of 1960.

As examples of the use of the two methods in localisation of ice cores in moraine ridges, the results of two parallell investigation from the Jotunheimen-measurements of October, 1959, will be shown here.

The first was a lateral moraine at Svellnosbreen, 30—50 meters high, which was thought to have an ice core. As can be seen from the seismic travel-time graph (fig. 2), the ridge only consisted of morainic material (sound velocity 710 m/sec.). Below this material the bed rock gives a velocity of 5,550 m/sec.

The earth resistivity curve from the same ridge (fig. 3) shows first a decrease of the earth resistivity, presumably because of increased humidity at greater depth. Thereafter it increases again, but not so rapidly as to indicate the existence of an ice core.

The second case, Veslegjuvbreen, has a terminal

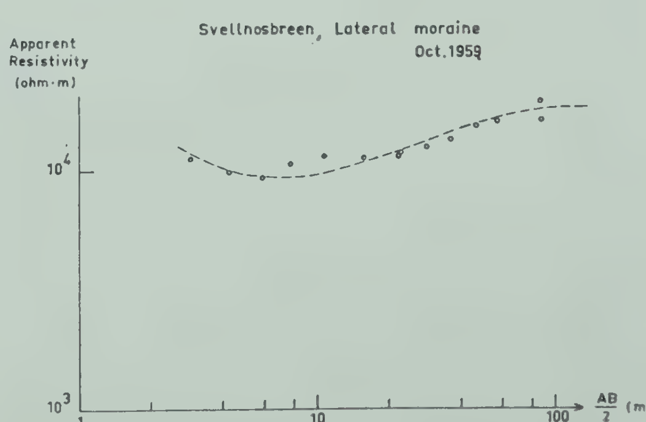


Fig. 3. Earth resistivity curve obtained on the lateral moraine ridges at Svellnosbreen. (The same locality as fig. 2, see text). Horizontal scale refers to the distance between current electrodes A and B.

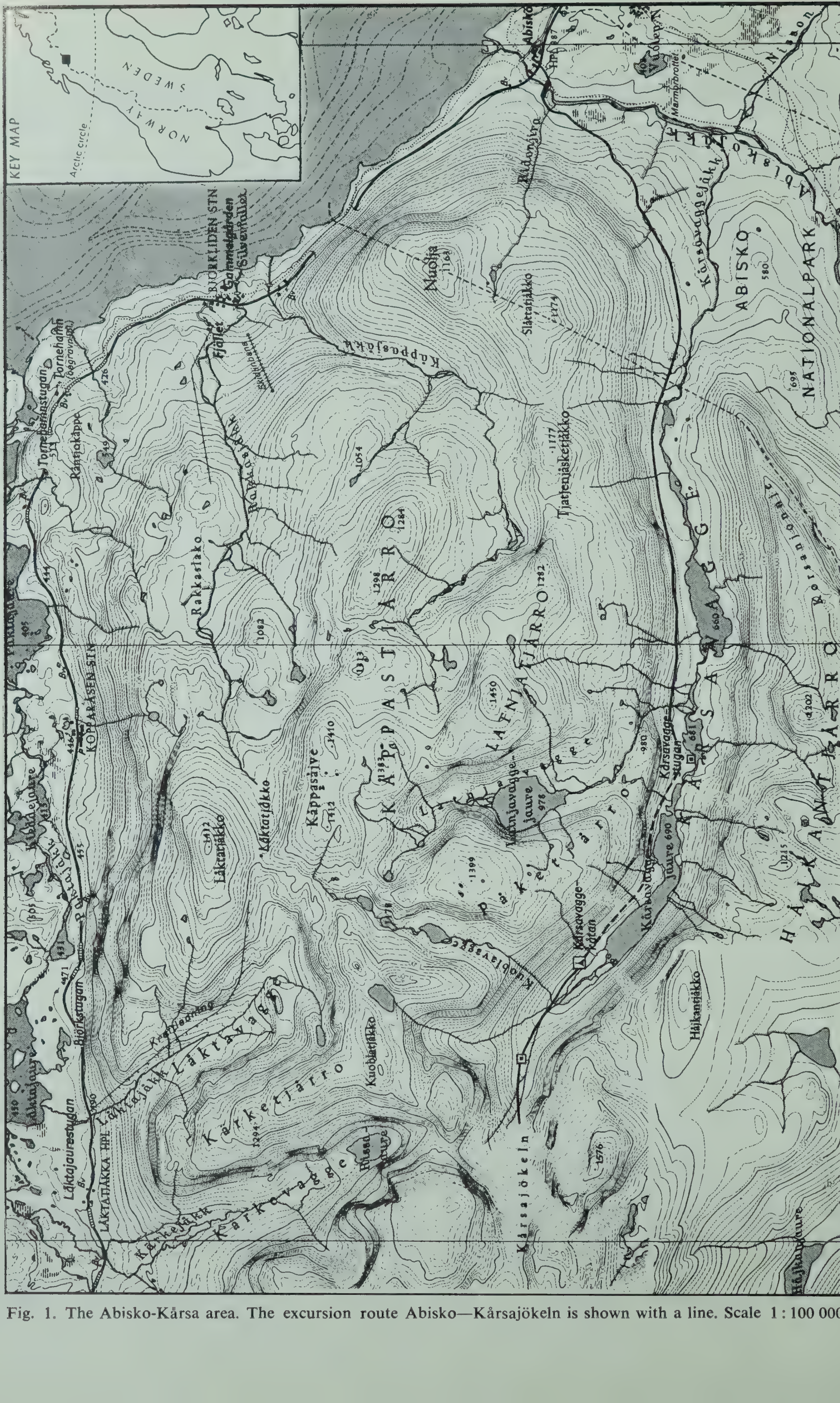


Fig. 1. The Abisko-Kårsa area. The excursion route Abisko—Kårsajökeln is shown with a line. Scale 1 : 100 000.

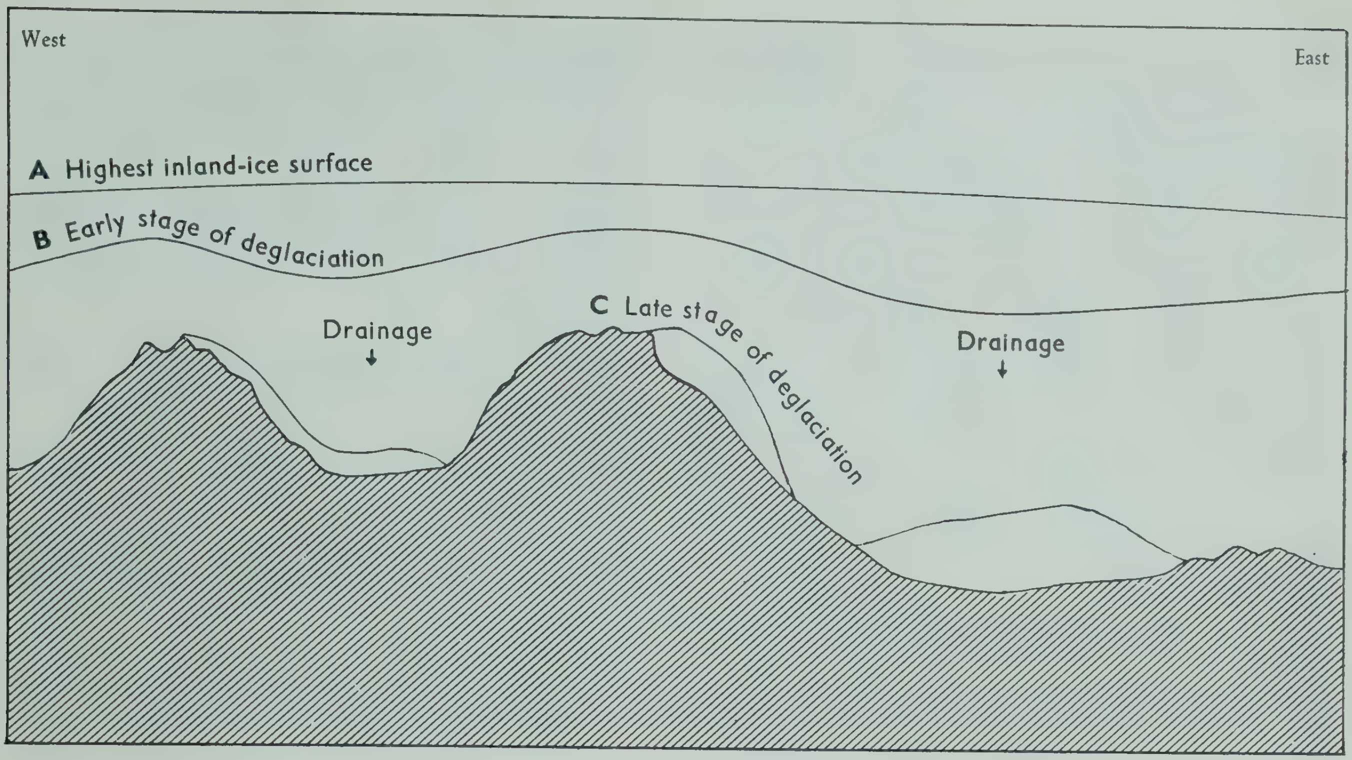


Fig. 2. Changes of ice surface. The highest level of the inland ice in northern Lapland was well above the highest summits (A). The main outlet valleys from the inland ice could be traced on the ice surface (B) soon after the beginning of the deglaciation stage. The latest stage of the deglaciation (C) in the Abisko-region was controlled by the topography and by the predominating west-wind drift of snow which was accumulated on the lee-sides.

When the inland ice declined, topography was soon reflected in its surface features. Fig. 2. Thinning of the inland ice meant that snow accumulation was insufficient, but for a long time the valleys in mountain districts delivered the same quantities of ice as before. Depletion went on through the valleys. The precipitation at high levels could not balance the discharge. In this way cupolas were established over high mountains long before any nunatakks were visible above the ice surface. These cupolas initiated glaciation centers during the deglaciation period in high mountain areas (Sarek, Kebnekaise, certain Norwegian mountain districts).

Eventually the glaciation centers from the initial period occurred again. They occurred in the same places as before. Some students have attached importance to erosion during the deglaciation stage, but, as has been shown, the recession was rapid, as it is in Alaska now, and the erosion was insignificant.

Terminal parts of the glacier tongues, which were far away from the centers of nourishment, were isolated. They died. *The Torneträsk ice* was such an isolated glacier body. It had one end near the present outlet of Torneträsk and the other end in the vicinity of the international boundary in the west. The highest point of this glacier body, which was about 80 km long from east to west, was about 1,100 m above sea level. Its culmination point was situated near Abisko. Thus, 700 meters of ice covered the Abisko valley approximately 8,000 years ago. Moraine and glacifluvial deposits all around the lake of Torneträsk had their distribution and morphology definitely fixed, directly and indirectly, by the Torneträsk ice. Its limitation horizontally and vertically is best indicated by lateral drainage channels.

Through striae observations and petrographic analysis of the till in the Abisko valley it is also possible to discover why the Torneträsk ice was thickest at Abisko. I have explained that the Abisko valley was one of the main drainage routes of the major glaciation center near Kebnekajse, and the two glacier tongues of the Torneträsk and Abisko valleys united at Abisko. This combination strengthened the glacier ice at the junction.

The terminal part of the glacier tongue in the Abisko valley is supposed to have been isolated from its snow-accumulation area. It was con-

sequently a glacier body during the final stage in the Abisko valley, but it remained attached to the Torneträsk ice like a lobe in spite of its genetic relations to another accumulation area. It should be kept in mind that the Abisko ice was long supplied from the west through the Kårsa valley.

Glaciologically the Kårsa area has a favourable position. The mountains all around the Kårsa Glacier are rich in snow, and the glaciation limit is well beneath the summits. Rock sculpture, striae, distribution of soil, and the general topography prove that the glacial conditions were the same at the beginning as at the end of a glaciation period. Probably there have not been any notable differences from one glaciation period to the other.

Imagine that we are about 50,000 years back in time. We have the same impression of the Kårsa Glacier as today, but from year to year it grows. The tongue protrudes into the Kårsavagge ("vagge" is the lapp word for valley). The glacier grows thicker and pours through the col northwards into the valley of Kärkevagge (1,190 metres above sea level) and subsequently through the col westwards into the valley of Vassivagge (1,260 m). Then the drainage tongues from the Kårsa glaciation center were developed.

Let us return to the glacier during deglaciation time. The tongues are retreating, but the scenery is not very different from that of the period when glaciers were growing at the beginning of the ice-age. We are especially interested in the glacier ice in the valley of Kårsavagge itself.

Morphology and deglaciation of Kårsavagge

In the summit region near Kårsavagge the ice moved towards the north, as is proved by striae. The higher up the striae are scratched, the older they are, generally speaking. And the striae which can be observed at the 1,500-metre level, for instance, in the mica-schist mountains north of Kårsavagge, should be ascribed to an early deglaciation period, when there were only few or no nunataks over the inland-ice surface.

This northerly direction left no traces at the bottom of the valley. Rock sculpture and rarely occurring striae indicate only one direction, i.e. towards the east. There has probably never been any other direction of ice movement in Kårsavagge.

The snow accumulation around the Kårsa Glacier was high enough to prevent an immediate collapse of the glacier tongue during the recession progress. There were small isolated dead-ice bodies which are easy to locate to-day through the pitted surface of till and of glacifluvial deposits. In places you can observe how brooks from the valley sides have taken part in the formation of the dead-ice topography by cutting the glacier ice to pieces and by transporting gravel and till on the receding ice, into it and underneath it.

Near the glacier hut a short tributary valley opens into the Kårsavagge. The concentration of dead-ice features at that point depends upon the junction of two glacier tongues, one in the Kårsavagge (the Kårsa Glacier) and the other in the tributary valley. As in many other places in the Kårsavagge, soil-creep has changed the forms but not very much.

Lateral drainage channels do not occur in Kårsavagge owing to the steep valley sides. Thus, one of the best indicators of glacier wastage and recession is lacking.

Terminal moraines provide other direct means by which the course of the ice recession has been disclosed in other parts of Sweden. Earlier students have interpreted accumulations in Kårsavagge as terminal moraines, but a detailed investigation has not proved the observations to be correct. There may possibly be one or two at the outlet of the highest lake in the valley. However, if they really are terminal moraines they do not necessarily prove that there was any change in climate during the latest stage of the ice-age. The deep basin of the lake inside the ridges, the shadowy position, and the vicinity of the cirque of the present Kårsa Glacier may have caused a stop of the glacier retreat, indicated by the low ridges.

The deglaciation of Kårsavagge is most suitably described from the mouth of the valley to the glacier.

The ice of Kårsavagge separated from the Abisko ice, when this was about to be isolated from its snow accumulation areas both in the south and southwest. The western accumulation area had been the Kårsavagge ice and its surroundings.

The melt-water from Kårsavagge deposited gravel and sand in the form of lateral terraces or deltas against the margin of the Abisko ice. Judging from these terrace fragments, the separation occurred at the 600-m level slightly inside the mouth of Kårsavagge. That is where the oldest terrace is situated. The Kårsavagge, like all other valleys in the Torneträsk district, is

hanging in relation to the Abisko valley, and on the slope, there is a kame-landscape, formed by eskers and remnants of terraces as well as transition forms between them, and kettle-holes are frequent.

The youngest in this series is at the same time the largest and the lowest. It is a terrace built up at about 410—420 m above sea level at the junction of three rivers, i.e. Abiskojokk, Nissonjokk, and Kårsavaggejokk ("jokk" pron. y o k k is the lapp word for river). This terrace is interesting only in connection with the Abisko ice, but a great deal of its material originates from the Kårsavagge.

West of the above-mentioned terrace at the 600 m level, glacifluvial material occurs for two or three kilometres, and it is even very free from till within a limited dead-ice area at the outlet of the easternmost lake (660 m above sea level). It is apparent that the shallow lake is dammed by the loose deposits.

In connexion with this glacifluvial area there is a small esker south of the river. The esker is aligned across the valley. Since it was deposited, there has been no movement in the glacier ice, so that the esker was destroyed. Therefore, the esker proves that the retreat was continuous and probably rapid, judging from the glacifluvial material in general. A rapidly-melting glacier produces much water, the capacity of melt water transport is great, and the till becomes transformed into glacifluvial material.

There is no clear limit to the glacifluvial zone of Kårsavagge. During a walk to the present Kårsa Glacier, however, you will eventually find that glacifluvial soil is no longer to be seen. Digging in the low hills and ridges produces no rounded particles. Everything looks like pure till. But most of the loose material has been transported by water, either by short streams or as soil creep. The particles might possibly have been rounded, were not the bedrock mainly schist and shale.

The morphological terminology of moraines is very vague, owing to the simple fact that moraine forms are mostly the result of movements in the glacier ice, where their development cannot be controlled. In Kårsavagge ridges at right angles to the valley-sides are common, and long-axis orientation indicates, in most cases, that the ridges are soil creep occurrences. You can often trace even the origin of a hill to a scar on the valley-side.

Activity in postglacial time is most clearly observed at the present glacier and partly in the highest lake (690 m above sea level).

The small terminal moraines in front of the glacier have been dated to the last two centuries by E. Bergström.

The delta in the lake is rapidly changing and filling the basin.

References

Atlas över Sverige. Map 21—22, Glacial striae, Terminal moraines, and Lines of ice recession; Map 23—24, Highest shore-lines of the sea and ice dammed lakes (authors: G. Lundqvist and E. Nilsson).

Holdar, C. G., 1952. Problemet Torne-issjön. Geogr. Annaler, 34.

— 1957. Deglaciationsförloppet i Torneträsk-området efter senaste nedisningsperioden, med vissa tillbakablickar och regionala jämförelser. Geol. För. Förh.,79.

THE KÅRSA GLACIER AND ITS RELATION TO THE CLIMATE OF THE TORNE TRÄSK REGION

BY C. C. WALLÉN
Department of Geography, University of Stockholm

Introduction

The Kårsa Glacier, situated in the Torne Träsk region of Swedish Lappland only 10 km from the Norwegian border and about as far south of the railway Kiruna—Narvik, is the most well-known of Swedish glaciers. Its size variations have been studied since the end of the last century and more regularly since 1909 (Svenonius 1910, Ahlmann and Tryselius 1929, Ahlmann and Lindblad 1940). In the nineteen-forties it was also thoroughly investigated as to the meteorological causes of these size variations (Wallén 1948—49). The glacier is situated in the isolated massif SW of Abisko of which the glacial-geological conditions are described by Holdar in the preceding article. It is a transition between a valley-glacier and a cirque-glacier probably regenerated during the deterioration of the post-glacial climate about 600 B.C., having had in the beginning and towards the end of the last ice-age a considerably larger extension.

The highest parts of the accumulation area of the actual glacier lie on a plateau of the Vuotasreita mountain at an elevation of about 1,500 m above sea level. From here the glacier slopes at first abruptly, later more gently, down in the Kårsavagge to its margin at about 800 m.s.l. The glacier has in its upper parts also a second contributory. It comes from the ridge between the Vassivagge valley in the west and Kårsavagge towards the east about 1,250 m.s.l. and slopes quite gently east-wards down into Kårsavagge, joining the first mentioned tongue at a height of 1,150 m and forming a widespread and quite important ablation area (Holdar fig. 1).

The most important part of the accumulation area is situated on the Vuotasreita plateau and on the slope from this plateau. The second part, beginning on the ridge between Vassivagge and Kårsavagge, is of little or no importance as an accumulation area. On the contrary the westernmost part of this tongue facing the Vassivagge valley is a secondary ablation area, and it is reasonable to agree with Ahlmann (1929) that the glacier "burns its light at both ends".

As is seen from the pictures (fig. 1) another glacier is situated on the slopes of the Vuotasreita mountain facing the Kårsavagge and the Kårsa glacier. This is called the Kårsa Side-glacier and since the beginning of the 1920ies has had no connection with the head-glacier. It therefore no longer has any importance in the "life" conditions of the Kårsa-Glacier.

Climate and glaciation in the Torne Träsk region

The Torne Träsk region of the Swedish mountain area is climatically a transition zone between an extremely maritime climate in the western part influenced by the Atlantic, and a more continental climate in the east. Generally speaking the region is maritime in character and shows an annual mean temperature of about 2° C above the normal for such latitudes within Sweden. December shows a positive temperature anomaly of 4° C, June a negative one of 3° C. The

Table 1. Monthly and annual means of temperature (°
Calculated temperature

	January		February		March		April		May		June	
	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.
Riksgränsen........	— 9.7	90	—10.5	61	—8.6	63	—4.1	60	+0.7	51	+6.2	70
Abisko............	—10.0	20	—10.8	15	—8.0	14	—3.0	11	+1.9	16	+7.4	30
Kårsa Glacier......	—		—		—		—		—		+2.0	

western parts of the region receive large amounts of precipitation but owing to topographical conditions the amount rapidly diminishes towards the east.

Table 1 shows the mean monthly precipitation and temperature at the meteorological stations Riksgränsen and Abisko located within the Torne Träsk region at an elevation of 500 m.s.l. and 320 m respectively. Temperature shows small differences between the localities, but there is a gradual increase in continentality towards the east, revealed by higher summer temperatures in Abisko than in Riksgränsen.

The mean annual precipitation in Riksgränsen, west of the massif where the Kårsa Glacier is situated, is 840 mm, while in Abisko the annual amount is only about 300 indicating the important drying up effect of the mountain barrier between the two places.

Temperature and precipitation determine to a large extent the glaciation conditions. The height of the glaciation limit in the westernmost part of the Scandinavian mountains is, according to Ahlmann (1924), around 1,000 m.s.l. In the western part of the Torne Träsk region the limit rises with the decrease of precipitation to around 1,300 m.s.l. Further east, around Abisko, it is already about 1,500 m.s.l. and rises rapidly eastwards. Precipitation is as indicated the decisive factor.

From the topographical map it will be seen that the local glaciation inside the area is surprisingly small; as a matter of fact the Kårsa Glacier is the only glacier of any importance. Ahlmann (1924) calculates the necessary annual amount of precipitation for glaciation in the western part of the Torne Träsk region to be 1,300—1,600 mm and states that the summer temperature (June—September) at the glaciation limit must not exceed + 1.6—+ 2.3°C. In table 1 the monthly temperatures of June—September are given for the Kårsa Glacier reduced with data from Riksgränsen to the period 1901—30 using measurements on the glacier carried out during the summers 1942—48. These data show that the summer temperature on the glacier has been somewhat above the limit given by Ahlmann, which is reasonable with regard to the retreat of the glacier that has occurred during the period.

It must be remembered also that local topographical features of a particular area can be of great importance in the creation of glaciers, especially in conjunction with prevailing winds. The Kårsa Glacier is a typical example of a case where glaciation has arisen in a valley where both the topographical and the meteorological conditions are particularly suitable. The innermost part of Kårsavagge serves as a kind of pocket where conditions regarding the prevailing westerly winds are very favourable for snow accumulation (cf. Holdar's discussion on the creation of the glacier in the beginning of the Ice-age).

The retreat of the Kårsa Glacier as related to the recent development of climate

The Kårsa Glacier, as most other glaciers in the northern hemisphere, has shown a remarkable retreat and thinning in recent years in conjunction with the considerably milder climate that prevailed during the first decades of this century.

Svenonius, who visited the glacier several times during the period 1884—1910, made observations of the position of the margin and found only small variations in the extension during this period. In 1909 a series of fixed points was created for continuous studies of the marginal variations. Nor were any important variations noted in 1917 when the stand of the margin was observed by means of these fixed points.

Since then, however, an accelerated retreat of the glacier has occurred. Wallén (1949) gives data for the variations in the stand of the northeastern part of the margin up to 1948. In table 2 additional data are given for the following years up to 1952 and for 1958. Unfortunately no measurements

d precipitation (mm) at Riksgränsen and Abisko (1901—30).
e glacier June—Sept.

	July		August		September		October		November		December		Year	
	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.	Temp.	Prec.
iksgränsen	+ 10.6	69	+9.4	66	+4.2	109	—1.6	64	—5.7	76	—9.2	64	—1.5	844
bisko	+ 11.1	42	+9.6	38	+5.2	28	—0.7	17	—5.4	17	—9.2	19	—1.0	267
årsa Glacier	+ 5.2		+4.2		+0.0		—		—		—		—	

were made in the intervening years. In 1958 a new series of fixed points were established because the distance to the old points is now so large that measurements from them are quite difficult. The complete picture of the variations of the stand of the margin is shown in table 3 and fig. 1 and 2.

It is evident that the retreat of the glacier has continued except for the year 1949—50 when a small advance was observed. In the period from 1952—1958 a quite remarkable retreat seems to have occurred, even though the summers were not particularly warm as will be seen from the table 4 showing mean monthly summer temperatures on the glacier calculated from Riksgränsen data reduced by means of coefficients obtained during investigations in the 1940ies (Wallén 1948). From these values the mean ablation season during the

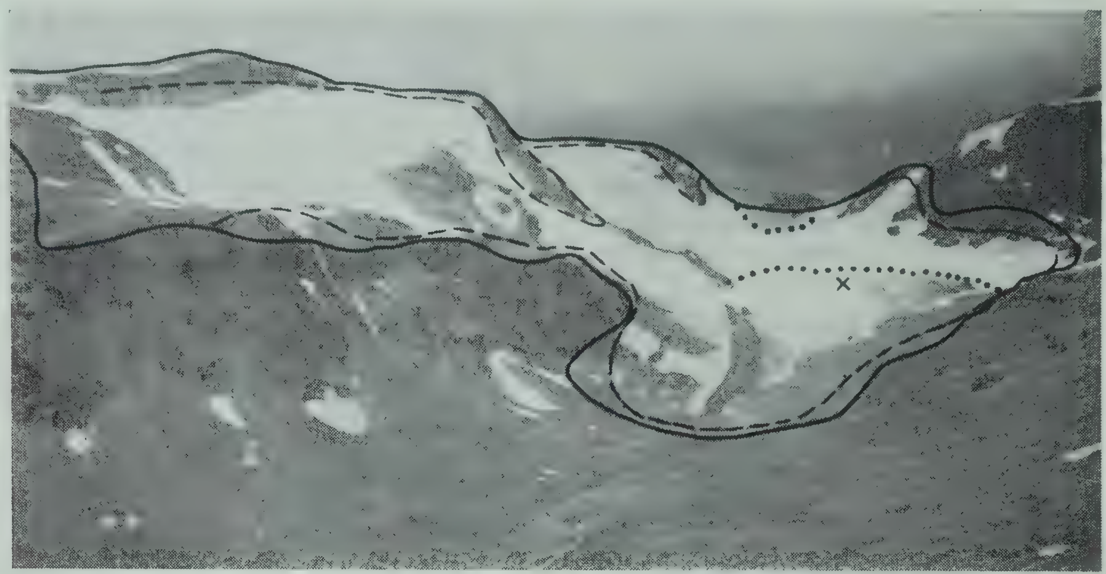


Fig. 1 a. The Kårsa Glacier photographed in 1948. Fulldrawn line indicates the glacier's extension in 1908, dotted line the extension in 1925.

Fig. 1 b. The Kårsa Glacier in 1959. Fulldrawn line indicates the stand of the tongue in 1948.

Table 2. Annual retreat of the Kårsa Glacier's margin in relation to Lindblad's fixed points of 1939.

Date of measurement	Retreat from the fixed points in m.									
	f 1	f 2	f 3	f 4	f 5	f 6	f 7	f 8	f 9	f 10
2.8.49	±0	±0	±0	±0	— 5.0	± 0	— 5.0	—2.5	± 0	± 0
3.9.50	±0	±0	±0	±0	+17.0	— 3.0	— 1.0	+3.5	+ 3.0	+ 1.0
2.8.51	±0	±0	±0	±0	—10.0	—10.0	—10.0	—6.0	—10.0	— 8.0
2.8.52	±0	±0	±0	±0	—28.5	—15.0	+ 1.0	—2.5	—14.5	—15.0
25.8.58	—	—	—	—	—	—	—	—	—88.0	—93.0

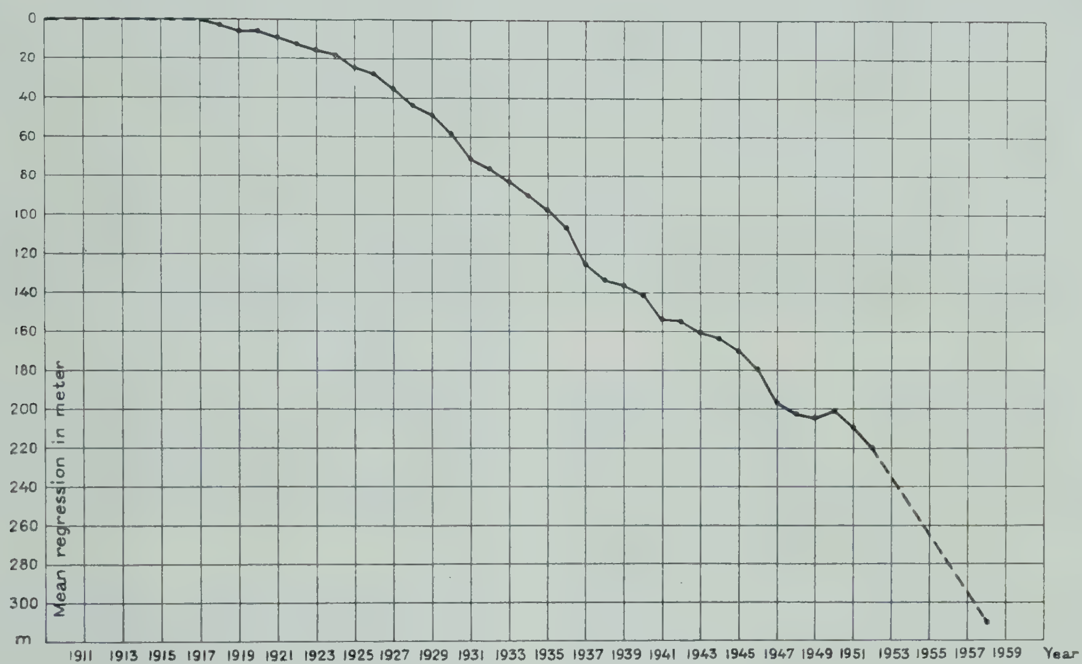


Fig. 2. Variation of the glacier's north-eastern margin since 1909 plotted additively.

period 1949—1958 has been determined to May 30—September 23, and the mean temperature during this season to + 3.4°. For 1930—39 the ablation season was calculated (Wallén 1948) to 25.5—20.9 and the mean temperature during this season to + 4.0°.

Table 3. Annual mean retreat of the glacier's north-eastern margin in m.

Year	Retreat	Year	Retreat
1909—17	± 0.0	1937—38	— 8.7
1917—19	— 2.7	1938—39	— 3.0
1919—20	± 0.0	1939—40	— 4.6
1920—24	— 2.9	1940—41	— 13.0
1924—25	— 6.0	1941—42	— 0.5
1925—26	— 3.5	1942—43	— 6.0
1926—27	— 7.5	1943—44	— 2.8
1927—28	— 8.5	1944—45	— 7.2
1928—29	— 5.0	1945—46	— 9.8
1929—30	— 10.2	1946—47	— 17.5
1930—31	— 12.7	1947—48	— 6.2
1931—32	— 5.0	1948—49	— 2.1
1932—35	— 6.8	1949—50	+ 3.4
1935—36	— 9.7	1950—51	— 9.0
1936—37	— 19.0	1951—52	— 10.8
		1952—58	(— 15.0)

The considerably lower summer temperature during the 1950ies points to an important change in the climatic trend during recent years. In the earlier investigation it was shown that a considerable and continuous increase of summer temperature played a dominant rôle in the rapid thinning of the Kårsa Glacier from around 1915 to the beginning of the 40ies. In fig. 3 overlapping means of temperature for ten years in Riksgränsen are given for winter and summer. In fig. 4 overlapping means of temperature for ten years are given for the different summer months.

First of all it is obvious that winter temperature shows a definite decrease during the last 20 years which has been the case all over Sweden. Maximum was reached in the decade 1929—38 with a mean temperature October—April of — 5.8°. In the last period, 1949—58 the mean temperature has decreased to — 7.3° i.e. a lowering of 1.5°. This temperature is comparable with those

Table 4. Calculated summer temperatures on the Kårsa Glacier 1949—58

	May	June	July	August	September
1949	—3.0	+2.8	+2.9	+2.7	+2.4
50	—4.0	+4.0	+6.3	+7.4	+2.0
51	—5.4	±0.0	+3.5	+7.0	+1.5
52	—3.5	+3.9	+4.6	+2.3	—0.2
53	—3.0	+6.5	+6.3	+5.2	+0.1
54	—0.5	+2.0	+7.4	+4.2	+1.3
55	—5.4	—0.1	+4.2	+4.1	+2.1
56	—2.4	+2.9	+7.0	+3.2	±0.0
57	—4.0	+1.0	+7.2	+3.5	+0.3
58	—4.1	+3.2	+3.9	+6.3	+1.4
Means	—3.5	+2.6	+5.3	+4.6	+1.1

recorded before the glacier retreat started. However, winter temperatures are usually of very little importance to the glacier's life except for those during the beginning and end of the accumulation season.

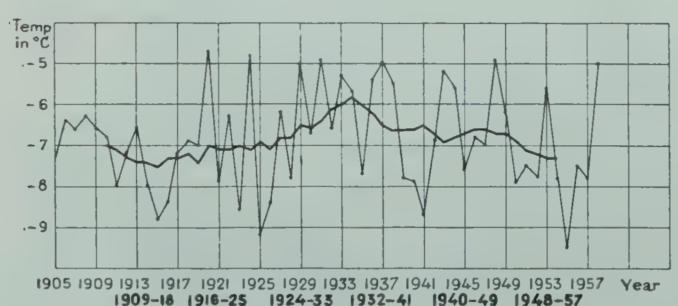
We therefore turn to the summer to look for changes in the trend. Even summer shows, as indicated, a typical cooling off. Maximum was reached during the decades 1930—39, 1932—41 and 1933—42, with mean temperatures May—September in Riksgränsen of + 7.3°. In the period 1949—58 it was only + 6.5° i.e. a lowering of 0.8° which corresponds very well to the lowering of the temperature of the ablation-season mentioned above.

If we look at the different summer months, we find that July shows the most considerable cooling off. The mean temperature during the maximum period 1930—39 was + 12.8°, while in July 1949—58 only + 11.5° was recorded. The curve moreover shows a continuous decrease. June and May indicate no important fluctuations but a change in trend from increase to decrease can be traced in June. August shows a particular picture in so far as temperature decreased from 1930—39 to 1940—49, but has then increased again so that the 1950ies have generally been warmer than the 40ies. This recent increase has of course considerably compensated the strong decrease during July as far as the temperature of the total ablation season is concerned.

September finally showed a continuous increase up to the period 1942—51 but has since become colder.

Looking at fluctuations in summer cloudiness we find from fig. 5 that a slight decrease has occurred since the 1930ies from 7.8 tenths to 7.4 on the average. This reduced cloudiness value points to a slight increase of incoming radiation during the 50ies compared with the 1930ies. There is no reason to believe that the radiation conditions have changed since the earlier period due to other causes than the decrease of cloudiness. Due regard will be taken to this decrease in the calculations below.

Unfortunately we face the same difficulties in estimating moisture and wind conditions during the 50ies as during periods used in the earlier



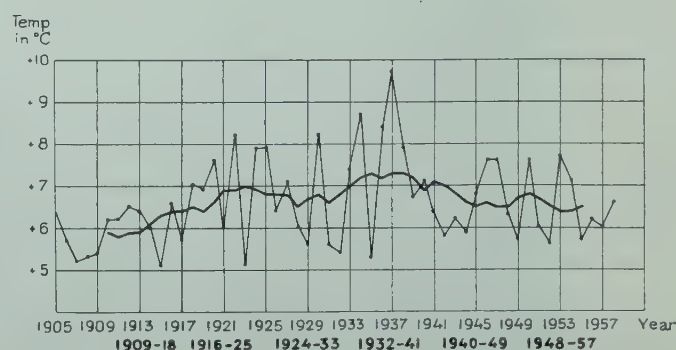


Fig. 3 a, b. Annual fluctuations and 10 years overlapping means of temperature at Riksgränsen in winter (October—April, a) and summer (May—September, b).

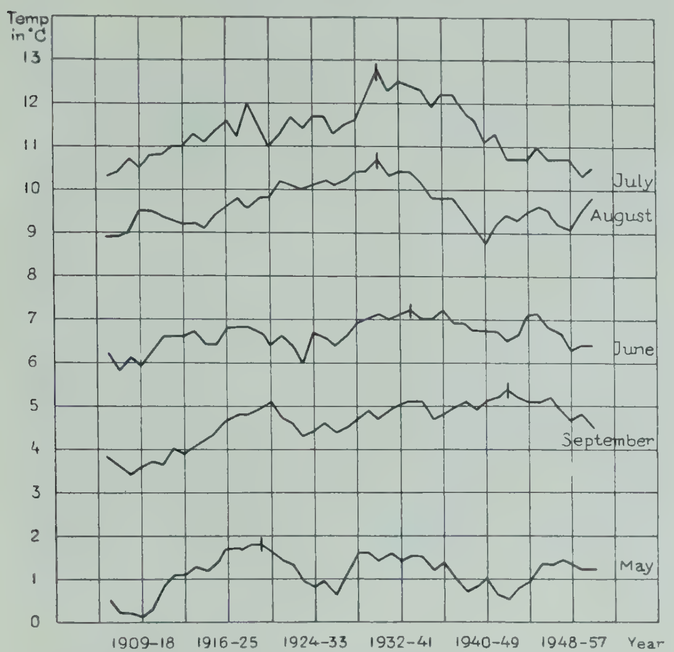


Fig. 4. 10 years overlapping means of temperature at Riksgränsen for different summer months.

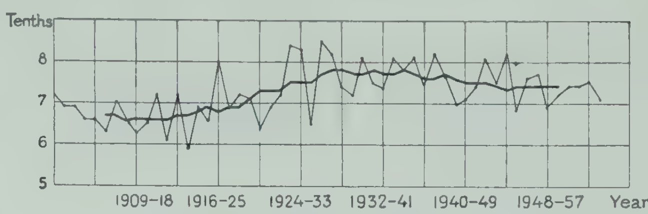


Fig. 5. 10 years overlapping means of summer cloudiness at Riksgränsen.

investigation. We have found no particular reason to assume any considerable changes in these elements from earlier periods and therefore have used the same values as during the 30ies in our further calculations.

It is now possible to calculate for the 1950ies the annual mean ablation on the Kårsa Glacier, using the same principles and formulae as were derived from results during the investigations in the 1940ies, of which it seems appropriate first to give a short summary here.

Importance of different meteorological factors in the ablation process

It was supposed from the beginning that the theory for exchange of momentum, heat and humidity above a snow-surface derived by Sverdrup (1936) from observations on Isachsens plateau should also be valid on the Kårsa Glacier. The analysis of data showed, however, that the exchange is much more complicated on a valley glacier of this type where local glacier breezes often produce a decrease rather than an increase of the wind-velocity above the surface. It was concluded that it was impossible to apply Sverdrups formulae to calculate the eddy conductivity over the snow-fields without special studies of the laws governing the exchange of momentum and heat in every particular case.

From records of temperature, humidity and wind-velocity at 170 cm above the surface we finally derived an empirical formula by which the transfer of heat from the air Q_a to the snow or ice surface could be calculated for any period of hours t

$$Q_a = 0.24 \left[\frac{1.8 \cdot 10^{-4}}{n_\Theta}\right] \cdot 170^{-\left(\frac{1}{n_\Theta}+\frac{1}{n_u}\right)} \cdot \Theta_{170} \cdot$$

$$\cdot u_{170} \cdot 3600 \cdot t \text{ or}$$

$$Q_a = \beta \cdot \Theta_{170} \cdot u_{170} \cdot 3600 \cdot t$$

where n_Θ and n_u are exponents determined by means of the variation with height of temperature Θ and wind-velocity u. An analogous expression was derived for the calculation of the amount of heat carried to the surface by condensation or away from the surface by evaporation.

From radiation and meteorological data received during different periods of several ablation seasons it was calculated that in the ablation of the snow cover the importance of radiation processes in spring is 60—75 % but decreases to a mean value of 37 % in late summer. The importance of the heat received from the air increases in the same period from 25 to 45 %. The heat supply through condensation at the surface is only 5 % in spring but rises to a maximum of some 20 % by midsummer and then decreases again slightly.

In the ablation on ice the importance of radiation is always much larger due to the smaller albedo of ice. Measurements of the albedo of snow and ice gave a mean value for an old wet snow surface of 60.3 % and of only 35.7 % for a wet ice surface.

It was also possible to determine which combination of weather conditions would be the most favorable for different kinds of ablation. First of all it should be emphasized that, owing to the maritime nature of the climate and the high moisture content of the air transported above, evaporation is a small or even insignificant factor in the total process of ablation on glaciers in northern Sweden. The most favourable

4—910100 Sep.

conditions for evaporation exist in spring with little cloudiness, high wind-velocity and low temperature. Under such conditions evaporation may be responsible for up to 10 % of the total snow ablation.

Favourable conditions for melting due to radiation were shown to be cloudless or partly clouded sky, low wind-velocity and low temperature. In spring such conditions may account for 100 % of the total ablation.

Melting due to heat received from the air reaches a maximum value with a half-covered to overcast sky, high wind-velocity and high temperature. These conditions gave in early summer the highest ablation per hour ever recorded on the Kårsa Glacier, namely 2.7 mm on snow and 3.5 mm on ice. In fact such conditions give up to 100 % larger ablation values than do the most favourable conditions due to radiation.

Calculated ablation amount during the 1950ies as compared with earlier conditions

Using data from Riksgränsen reduced with values derived during the investigations in the 40ies, and applied in the above mentioned empirical formula, the ablation due to different factors was calculated for the two periods 1906—15 and 1930—39 of which the former was comparatively cold and the latter instead quite warm.

Our calculations for the 1950ies have now been made in an analogous way and are compared with the results from the earlier periods in table 5.

As mentioned above the calculated length of the ablation season during the 1950ies is 116 days between May 31 and September 23. Using the average cloudiness value of 7.4 we obtain a value of 51,000 cal/cm^2 for the average incoming radiation during the ablation season in the 1950ies. The outgoing radiation also changes with the length of the season and the cloudiness. It was calculated to amount to 9,700 during the 1950ies average ablation season.

Assuming that our reference point at 1,025 m.s.l. (the camp level during the 1940 investigations) on the average has been below the firn limit after July 1 we have to put the albedo in the first period of the ablation season at 60 % and during the latter at 37 %. By means of these values we get, with monthly values of incoming and outgoing radiation, a total amount of energy received from radiation processes of 13,000 cal/cm^2.

Using the interpolation formula mentioned earlier we have calculated the convectional and condensational heat amounts used for ablation applying a mean temperature of $+3.4°$, an average wind-velocity of 2.3, and a mean vapour pressure of 5.2 mm/Hg. With these values we obtain

$$\beta = 18 \cdot 10^{-4}$$

and consequently

$$Q_a = 18 \cdot 10^{-4} \cdot 3.4 \cdot 230 \cdot 2784 \simeq 3900 \text{ cal/cm}^2$$

$$600F = 18 \cdot 10^{-4} \cdot 0.6 \cdot 230 \cdot 2784 \simeq 1600 \text{ cal/cm}^2$$

The mean ablation calculated in this way corresponds to 163 cm carried away through radiation energy, 49 cm through convection and 20 cm through condensation or altogether 232 cm per year indicating a decrease since the 1930ies of 6 % in the annual mean ablation (table 5). This means that the amount melted by radiation

Table 5. Calculated annual mean ablation at the camp level on the Kårsa Glacier during the periods 1906—15, 1930—39 and 1949—58.

	$\alpha I - R$	Q_a	600 F	Melt by radiation	Melt by convection	Melt by condensation	Total
	cal/cm^2	cal/cm^2	cal/cm^2	cm of water	cm of water	cm of water	cm of water
1906—15 (Snow)	8,900	3,500	1,500	111	44	19	174
1930—39 Snow, Ice	12,000	4,900	2,100	161	61	26	248
1949—58 Snow, Ice	13,000	3,900	1,600	163	49	20	232

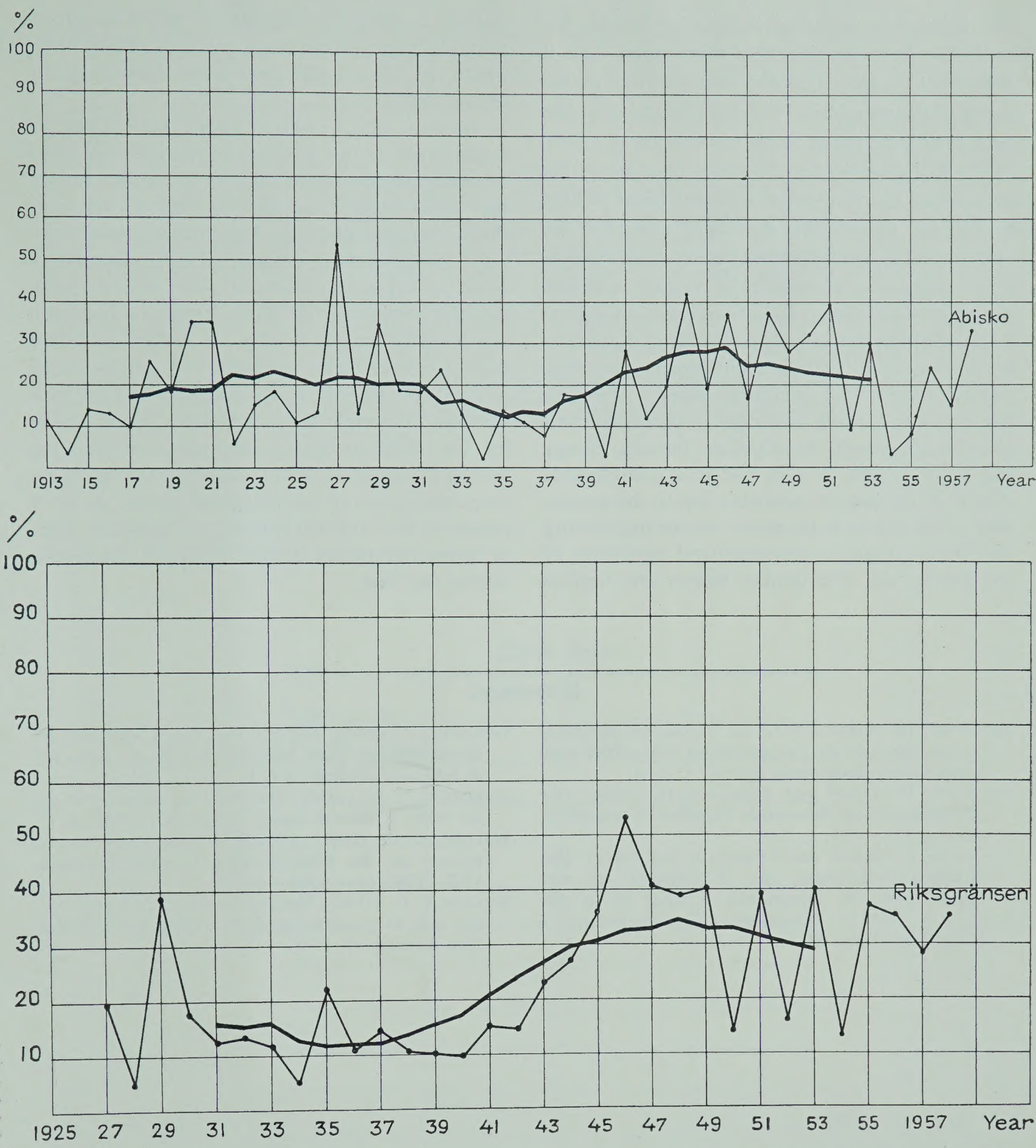


Fig. 6. Annual fluctuations and 10 years overlapping means of percentage of rainfall in relation to total precipitation October—May at Abisko and Riksgränsen.

was during the 50ies 70 % of the total or 5 % higher than during the 30ies while the convectional and condensational part of the ablation had diminished by 4 and 1 % respectively. Although the changes are not great they indicate a slight return to earlier conditions, when, if we assume both snow and ice ablation to have occurred at the reference point, the radiation part of the ablation must have been larger than during the 30ies. This of course must be due to the slight decrease in temperature and length of the ablation season.

Although the mean ablation during the 50ies must have been only slightly smaller than during

the 30ies the retreat of the glacier margin has probably been considerably more rapid. This is difficult to verify but the measurement of the stand of the margin in 1958 shown in fig. 1 points to a very large retrat of the glaciers margin since 1952. Not considering the slight changes in the percentage distribution of different kinds of heat supply for the glacier, the rapid retreat in the last six years must have been due to similar causes that created the considerable retreat in earlier decades. These causes have been concluded to be a considerable heat supply conveyed to the glacier by convectional and condensational processes during a quite long ablation season. Although the ablation season has slightly decreased and also the heat supply from the air, the total energy has been enough to give rise to a considerable retreat of the glacier, probably due to the smaller area of the glacier in the recent decade than during the 30ies and also to topographical conditions of the glacier bed. The smaller glacier area implies namely a comparatively greater effect of advectional influences from the snow-free surroundings, which are not taken into consideration in the above calculations.

Another effect which might have been of importance is the increase of number of cases of temperatures above the freezing point during the accumulation season. In fig. 6 we have shown the percentage of rainfall in relation to the total precipitation amount during October—April at Riksgränsen and Abisko. Fig. 6 shows that the percentage of rainfall amount has been considerably greater during the 1940ies and 50ies than it was in earlier periods. The general trend is given by 10-year overlapping means at both stations. Through the calculation given above for the ablation season the effect of thawing during the accumulation season cannot be taken into consideration but the trend shown in fig. 6 points to the fact that this effect might have been of some importance in the retreat of the glacier during the 50ies.

References

AHLMANN, H. W:SON (1924): Le niveau de glaciation comme fonction de l'accumulation d'humidité sous forme solide. Geografiska Annaler 6: 3—4.

AHLMANN, H. W:SON and TRYSELIUS, O. (1929): Der Kårsagletscher in Schwedish Lappland. Geografiska Annaler 11: p. 313.

AHLMANN, H. W:SON and LINDBLAD, TH. (1949): Die Grössenveränderungen des Kårsajökels in den Jahren 1909—39. Geografiska Annaler 22; p. 80.

SCHYTT, V. (1947): Glaciologiska undersökningar i Kebnekajse. Ymer 67: 1.

SVENONIUS, F. (1910): Studien über den Kårso- und die Kebnegletscher nebst Notizen über anderer Gletscher in Jukkasjärvigebirge. S.G.U. Ser. Ca No 5.

SVERDRUP, N. U. (1936): The eddy conductivity of the air over a smooth snow field. Geof. Publ. 11: 7.

WALLÉN, C. C. (1948): Glacial-meteorological investigations on the Kårsa-Glacier, Swedish Lappland, 1942—1948. Geografiska Annaler 30: 3—4.

WALLÉN, C. C. (1949): The shrinkage of the Kårsa Glacier and its probable meteorological causes. Geografiska Annaler 31, 1—4.

Date Due

~~SEP 30 '74~~			
OCT 7 '74			

ESSELTE, STHLM 60